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The role of thermal convection in heat and mass transport in the subarctic snow cover

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THE ROLE OF THERMAL CONVECTION IN HEAT AND MASS TRANSPORT

IN THE

SUBARCTIC SNOW COVER

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THE ROLE OF THERMAL CONVECTION IN HEAT AND MASS TRANSPORT

IN THE

SUBARCTIC SNOW COVER

A

THESIS

Presented to the Faculty of the University of Alaska

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DOCTOR OF PHILOSOPHY

By

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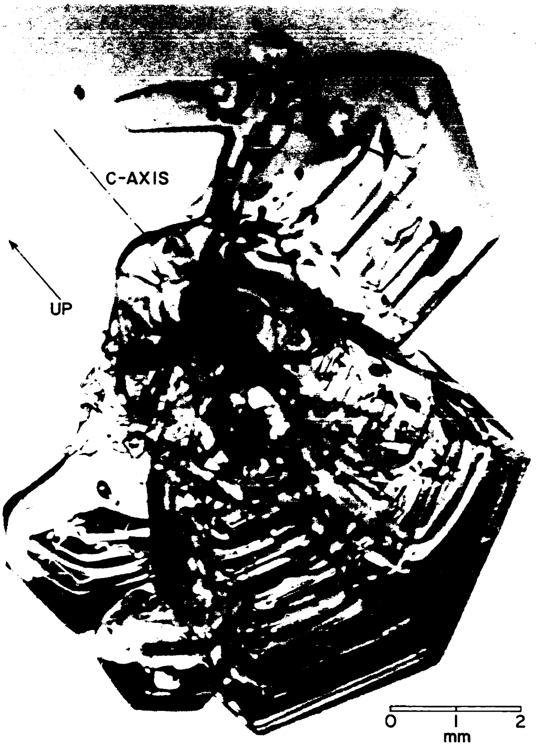
FRONTISPIECE: A depth hoar crystal from the Fairbanks snow cover collected January, 1987, magnified 18X. The crystal is a pyrimidal cup from near the base of the snow (see Section 3.1.3). The sharp lower edges indicate the crystal was actively growing at the bottom, and the rounded upper edges indicate it was eroding by sublimation at the top. Striae and steps on prism faces are typical of the depth hoar found in the subarctic snow cover.


C-AXIS
UP
0
1
2
mm

ABSTRACT

The purpose of this study was to investigate the role of air convection in moving heat and water vapor in snow. To detect convection, the three dimensional temperature field in the Fairbanks snow cover was measured hourly during three winters (1984-1987). Measurements of snow density, compaction, and grain size were made monthly to determine the vapor flux and textural changes.

The snow metamorphosed into depth hoar, producing a sequence of five layers, including a basal layer with horizontal c-axes. C-axes in the overlying layers were vertical or random. As the depth hoar developed, its air permeability increased to a value several times higher than previously measured for any snow, while the number of snow grains per unit volume decreased by an order of magnitude as a few select grains grew while others sublimated away. Simultaneously, there was a net transfer of mass from the base to top of the snow due to mass flux gradients that averaged 3×10^{-6} kg $m^{-2}s^{-1}m^{-1}$, but were occasionally 10 times higher.

Convection occurred sporadically in 1984-85 and continuously in 1985-86 and 1986-87. The evidence was 1) simultaneous warming and cooling at different locations in a horizontal plane in the snow, and 2) horizontal temperature gradients of up to 16 K m^{-1}. The convection was time-dependent, with perturbations such as high wind or rapid changes in air temperature triggering periods when horizontal temperature gradients were strongest, suggesting these were also periods when the air flow was fastest. During the winter, warm and cold zones developed in the snow and remained relatively fixed in space. The zones were probably the result of a diffuse plume-like convection pattern linked to spatial variations in the temperature of the snow-soil interface. Air flow was inferred to have been horizontal near the base of the snow and vertical elsewhere. Flow

averaged 0.2 mm s^{-1}, with a maximum of 2 mm s^{-1}. During average flow conditions, convection moved about a third of the total heat, but did not move significant mass. However, the coincidence of crystals with horizontal c-axes and the horizontal flow lines at the base of the snow suggests that convection may have affected crystal growth direction.

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A Note on Units

Units and symbols are consistent with the SI system.

A	coefficient of exponential relationship
a_1, a_2, etc.	coefficients of polynomial fit to data
B	exponent of exponential relationship
b_1, b_2, etc.	intercepts of linear fit to data
c	specific heat capacity [J kg^{-1} K^{-1}]
c_t	specific heat capacity of thermistor [J kg^{-1} K^{-1}]
$\bar{d}$	mean "grain size" from sieving or photography [mm]
$\bar{D}_{j,j+1}$	central sieve mesh diagonal [mm]
D_o	diffusion coefficient of water vapor in air [$m^2 s^{-1}$]
e	total uncertainty in calculated result
e_h	uncertainty in measurement of layer thickness [m]
e_{Dh}	uncertainty in change in layer thickness [m]
e_ρ	uncertainty in measurement of density [kg m^{-3}]
$e_{D\rho}$	uncertainty in change in density [kg m^{-3}]
$\dot{e}$	strain rate [s^{-1}]
$\dot{e}_{zz}$	vertical strain rate [s^{-1}]
F	enhancement factor

g	acceleration of gravity [m s^{-2}]
h	total snow depth or total snow layer thickness [m]
h_b	basal snow layer thickness [m]
H	geometric parameter of HFM (Heat Flow Meter)
i	electrical current [amp]
J	mass flux per unit area per unit time [kg $s^{-1}m^{-2}$]
$\partial J/\partial z$	vertical mass flux gradient [kg $s^{-1}m^{-2}m^{-1}$]
$\nabla \cdot \bar{J}$	mass flux divergence [kg $s^{-1}m^{-2}m^{-1}$]
k	thermal conductivity of snow [W $m^{-2}K^{-1}$]
k_{dry}	thermal conductivity w/out latent heat transfer [W $m^{-2}K^{-1}$]
k_{eff}	bulk effective thermal conductivity including heat transport by convection [W $m^{-2}K^{-1}$]
k_m	thermal conductivity of HFM [W $m^{-2}K^{-1}$]
k_s	thermal conductivity of material around HFM [W $m^{-2}K^{-1}$]
L	number of sieves, also depth below snow surface
L_s	latent heat of sublimation [J kg^{-1}]
m	mass [kg]
$\bar{m}_j$	average mass of a grain in the j^{th} sieve [kg]
m_t	mass of thermistor [kg]
M_j	weight fraction in the j^{th} sieve [kg]
n	number of observations
N	number of grains per unit volume
N_j	total number of grains in the j^{th} sieve

N_T	total number of grains in a sample
Nu	Nussult number
P	electrical power [W]
Pe_h	Peclet number for heat
Q_m	heat flow measured by heat flow meter [J $m^{-2}s^{-1}$]
Q_s	heat flow at the snow-soil interface [J $m^{-2}s^{-1}$]
r	correlation coefficient, also radial coordinate
R	radius of cylindrical hole
Ra	Rayleigh number
Ra_{cr}	critical Rayleigh number (onset of convection)
t	time [s]
T	temperature [°C]
ΔT	change in temperature [K]
$\partial T/\partial z$	temperature gradient [K m^{-1}]
$\partial^2 T/\partial z^2$	curvature in vertical temperature profile, [K m^{-2}]
$\bar{v}$	air flow velocity [m s^{-1}]
v_z	vertical air flow velocity [m s^{-1}]
V	volume [m^3]
ΔV	small control volume [m^3]
W	watts
z	vertical coordinate [m]
z_t	thermistor height above ground [m]

Greek Symbols

α	thermal diffusivity [m^2 s^{-1}]
α_{air}	thermal diffusivity of air [m^2 s^{-1}]
α_{ice}	thermal diffusivity of ice [m^2 s^{-1}]
α_{snow}	thermal diffusivity of snow [m^2 s^{-1}]
α_{dry}	thermal diffusivity without latent heat transfer [m^2 s^{-1}]
β	isobaric coefficient of thermal expansion [K^{-1}]
κ_i	intrinsic permeability [m^2 s^{-1}/N m^{-2}]
ν	kinematic viscosity [m^2 s^{-1}]
ρ	bulk snow density or bulk density of a snow layer [kg m^{-3}]
ρ_v	water vapor density [kg m^{-3}]
$\partial\rho/\partial t$	bulk densification rate including compaction [kg $m^{-3}s^{-1}$]
$\partial\rho/\partial t]_v$	densification from vapor flux only [kg $m^{-3}s^{-1}$]
ρc	volumetric heat capacity [J m^{-3} K^{-1}]
$(\rho c)_f$	volumetric heat capacity of air [J m^{-3} K^{-1}]
$(\rho c)_m$	volumetric heat capacity of porous medium [J m^{-3} K^{-1}]
$(\rho c)_s$	volumetric heat capacity of ice [J m^{-3} K^{-1}]
τ	response time of snow to temperature change [s]
ϕ	porosity, or grain size measure if associated with sieving
Ω	electrical resistance [ohms]

ACKNOWLEDGEMENTS

"To see a World in a Grain of Sand..."

William Blake

I had always seen myself doing science projects where I skied over miles of icecap and glacier, rather than ones where I spent hours lying in the snow looking at crystals with a hand lens, or days thinking about vapor molecules moving a few millimeters. Many people have helped me to see this fascinating other world and to mature as a scientist.

Dr. Carl Benson, my chairman, has been a friend, inspiration, exasperation, and delightful mentor for these past eight years. Thirty years of working in the snow has not dampened his enthusiasm; I hope I can say the same thirty years from now.

Dr. Jerry Johnson, also a member of my committee, initiated this project, rolled soil and strung thermistor wires, listened as my physical interpretations went through radical changes, supplied physical insight when I needed it, and helped insulate me from the demands of work in order to have the time to write up this dissertation.

Drs. Sue Ann Bowling, Lew Shapiro and Will Harrison have all contributed, not just to this dissertation, but to my education as a scientist, both in and out of the classroom.

Many other people contributed their time, thought and labor to this project and deserve thanks and praise. Ed Chacho coordinated his permeability measurements with my work, helped dig snow pits and provided many good ideas. Dr. Don Perovich worked on analyzing the data from 1985, came up with the correlation analysis which appears in this dissertation, and provided good, skeptical comments at the right time. Steve Perkins wrote computer operating and analysis programs which were critical to the success of the work, given the magnitude of the data set. Monique Fouchet and Jim Morse fabricated many of the thermistor strings. Ma.: Stark, Joe Holty and Larry Burke at the University of Alaska farm helped in many ways.

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1 INTRODUCTION

1.1 Statement of Purpose

The purpose of this study was to determine if air convected in a natural snow cover, and if it did, to evaluate its importance in moving heat and water vapor through the snow. To do this, measurements of temperature and density, from which heat and mass flow could be calculated, were made in the dry, cold snow cover near Fairbanks, Alaska.

The question of whether convection occurs in natural snow covers has been debated for decades. Seasonal snow covers lie on ground which is generally warmer than the air. This condition creates an unstable density stratification in the air in the snow which is necessary, but may not be sufficient, to cause thermal convection. Early investigators suspected that convection occurred, basing their speculation on inconclusive field tests and the failure of their diffusive mass transport models to predict the observed mass flux in the snow (Bader et al., 1939). More recently, attempts to induce convection in laboratory experiments have had mixed results, with convection occurring in some experiments (Powers et al., 1985a; 1985b) but not in others (Akitaya, 1974). Theoretical studies have been equally inconclusive, suggesting that convection was likely to occur only in deep, highly permeable snow covers (Palm and Tveitereid, 1979; Powers et al., 1985a; 1985b).

The only germane field studies were done by Trabant and Benson (1972) and Gjessing (1977). These authors concluded from their measurements that convection occurs in natural snow covers, however the presence of convection was inferred from indirect evidence in both cases. For example, the former authors based their conclusion on the fact that the measured mass transport in the Fairbanks snow cover was too great to be accounted for by existing theories of diffusive mass transport. However, recent diffusion models (Sommerfeld, 1983; Colbeck, 1983a; Gubler, 1985) could account for the greater transport. The present study was undertaken because the existing evidence for convection in the snow, both theoretical and experimental, was contradictory and inconclusive.

The subarctic snow found near Fairbanks, with its extreme temperature gradients and permeability (Figs. 1-1, 1-2), appeared to be the ideal place to make the measurements necessary to determine if convection occurs in natural snow covers. Though of moderate depth (Fig. 1-3), the Fairbanks snow cover is subjected to temperature gradients many times stronger than those found in other snow covers (Fig. 1-1), generating highly unstable density stratification of the air in the snow. The snow completely metamorphoses into depth hoar during the winter. As this metamorphism proceeds, the air permeability increases, becoming 2 to 3 times more permeable than any previously measured snow (Fig. 1-2 and Appendix I), thus increasing the likelihood of convection. Also, a large body of observations made by Benson (unpublished), Trabant (1970), Benson and Trabant (1972), and Trabant and Benson (1972) at the same experimental site used in

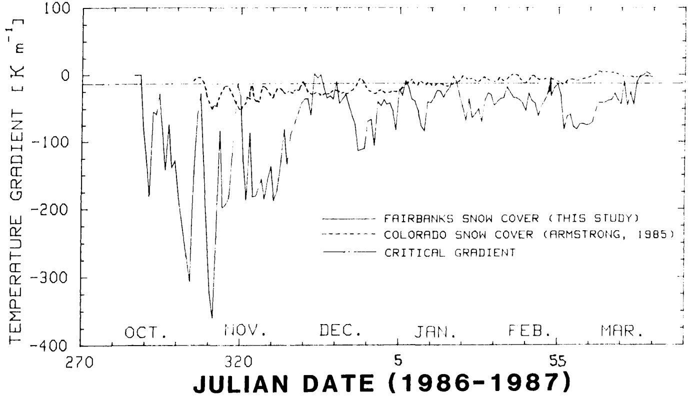

FIGURE 1-1: The average vertical temperature gradient across the Fairbanks snow cover during 1986-87. For comparison, the temperature gradient for a snow cover in Colorado is also shown (Armstrong, 1985). The critical temperature gradient necessary for the formation of depth hoar (see Section 1.3.2) is approximately 10 K m^{-1}, shown by a dot-dash line. Temperature gradients in the Fairbanks snow cover exceed the critical gradient for more than 150 days.

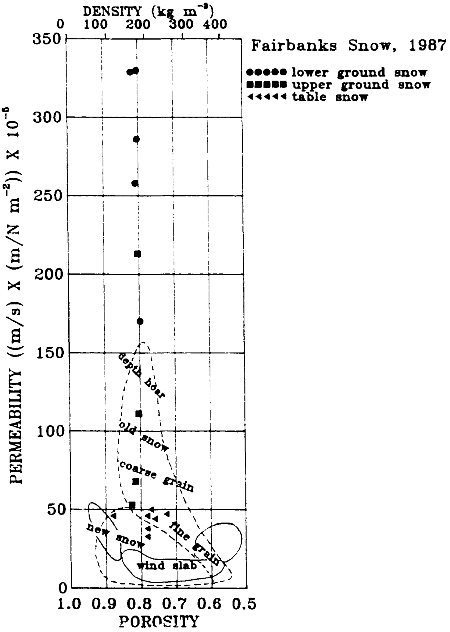

FIGURE 1-2: The air permeability of the Fairbanks snow. Other snow types are shown by fields enclosed by solid lines (Shimizu, 1970) and dashed lines (Bader et al., 1939). The permeability of Fairbanks snow (ground) is 2 to 3 times greater than previously measured depth hoar, and 5 to 7 times more permeable than snow not subjected to strong temperature gradients because it was deposited on a table.

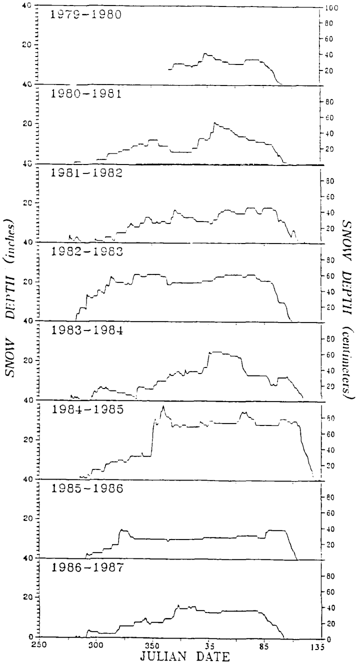

FIGURE 1-3: Snow depth in Fairbanks, 1980 to 1987. Data from the Fairbanks International Airport collected by the National Weather Service.

the present study suggested that convection occurred in the Fairbanks snow cover.

Intensive measurements were made in the Fairbanks snow during the winters of 1984-1985, 1985-1986, and 1986-1987 (called here the winters of 1985, 1986, and 1987 respectively). Because the movement of air could not be measured directly, natural thermal convection was identified by its perturbation of the three-dimensional temperature field in the snow, which was measured using an array of thermistors. To evaluate the role of convection in mass transport, precise density and grain size measurements were made in order to determine the movement of water vapor and the metamorphic changes in the snow.

The vapor movement could not be ignored because heat and mass transport are inextricably linked in the snow cover. Ice, which is generally near its melting point, has a vapor pressure which is several orders of magnitude higher than that of most other geologic materials. The vapor pressure is also an exponential function of temperature (Fig. 1-4 and Section 5.3.3). Consequently, the strong temperature gradients in the Fairbanks snow cover produce vapor pressure gradients which cause sublimation from warm grains and deposition on cold grains, transferring both latent heat and mass. The temperature gradients in the Fairbanks snow are not only stronger (Fig. 1-1), but also include higher temperatures than many other snow covers. Thus, the absolute values of the vapor pressure are higher. For example, with an snow surface temperature of -45°C, the base of the Fairbanks snow cover might be about -10°C. At equal depth below the surface on the Greenland Ice Sheet, the temperature would be

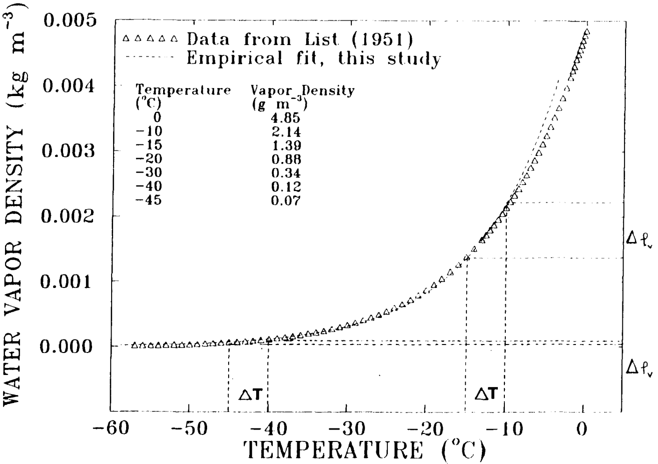

FIGURE 1-4: The saturation water vapor density over ice . The data (▵) have been fit with an exponential curve (-----) $\rho_v = A e^{BT}$, where A = 0.005789, B = 0.09658 for T in °C (see Section 5.3). Temperture increments of 5° between -10° and -15°, and -40° and -45° produce vapor density changes that differ by a factor of 17 (stippled areas).

about -40°C. An increment of 5° between -10° and -15°C corresponds with a change in water vapor density of 0.75 X 10^{-3}kg m^{-3}; a 5° increment between -40° and -45°C corresponds with a water vapor density change of 0.05 X 10^{-3}kg m^{-3}. These differ by a factor of 17 (Fig. 1-4). (Vapor density rather than vapor pressure is used in this study because it is more convenient in calculations. It can be converted to the vapor pressure using the ideal gas law). In Fairbanks, the 5° increment would occur over a vertical distance of 0.07 m, but would require a distance of about 0.5 meters in the deep snow cover on an ice sheet such as Greenland, so the vapor density gradient in the Fairbanks snow would be 100 times stronger.

The mass transport also results in snow metamorphism. In Fairbanks, the end result of the metamorphism is the transformation of the entire snow cover into depth hoar. It was this striking texture that led Benson and Trabant (1972) to investigate the mass flux in the first place. One goal of this investigation was to determine if convection played an important role in the development of depth hoar, and conversely, the role played by depth hoar in establishing and maintaining convection. Depth hoar has been the subject of scientific and engineering interest for many years because of its role in providing a failure plane for avalanches.

There are several reasons why it is important to know if air convects in the snow. If convection occurs it will increase the effective thermal conductivity of snow, which in turn means that the ground under snow in which air is convecting would freeze deeper during the winter than ground under snow in which there was no convection. This is important to agriculture. The survival of under-snow biota also depends on how well the snow insulates them from ambient air temperature (Pruitt, 1984). Convection may move contaminants into the snow (Gjessing, 1977) and affect the mixing of isotopes (Friedman et al., in prep.). Environmental concerns make the former important, and the latter has a direct impact on the interpretation of snow pits and ice cores in polar regions, where depth hoar layers form the primary annual stratigraphic marker in ice sheets and glaciers.

Though the goal of this study was to determine if convection occurred in the snow, several related questions were of interest:

1) If convection occurs, what flow pattern develops (see Section 5.2.1)?
2) What are the relative contributions of conduction, vapor movement with latent heat transfer, and convection of air to the overall heat transfer in snow (see Section 5.4)?
3) Can specific metamorphic changes in the snow be attributed to convection (see Section 5.4)?
4) Do existing theories adequately describe vapor transport in snow (see Sections 1.1 and 1.2)?

1.2 Background

Key points from the literature in heat and mass transport in snow are summarized below. Several relevant studies on convection in porous media other than snow are discussed in Section 5.2.

1.2.1 Diffusive Mass Transport in Snow

Paulke (1934a, 1934b) and Seligman (1936) determined from their field observations that temperature gradients in the snow resulted in water vapor density gradients which drove vapor diffusion. Subsequently, at least 7 one-dimensional mathematical models of mass transport by water vapor diffusion have been developed (Bader et al., 1939; Yosida et al., 1955; Giddings and LaChapelle, 1962; Yen, 1963; de Quervain, 1972; Palm and Tveitereid, 1979; Fedoseeva and Fedoseev, 1988). The diffusive mass transport predicted by the model of Bader et al.(1939) was so small it led the authors to conclude that convection must transport considerable mass. When their model results and observations did not agree, other authors (Yosida et al., 1955; Yen, 1963; de Quervain, 1972; Palm and Tveitereid, 1979; Fedoseeva and Fedoseev, 1988) concluded that the diffusion coefficient cf water vapor was enhanced in the snow cover.

Yosida et al. (1955) used an experimental apparatus to measure the actual vapor flux gradient in a snow layer. Comparing model results to measurements, they determined that an enhancement factor (F) of approximately 5 was necessary for the model to produce the measured mass transport. Yen (1963), de Quervain (1972), Palm and Tveitereid (1979) and Fedoseeva and Fedoseev (1988), in comparing theoretical and experimental results, also found that they needed enhancement factors of approximately 5 to bring their calculated results into agreement with measurements.

By assuming the diffusion coefficient was enhanced (or there were enhanced temperature gradients) to bring model results in line with measured values of mass transport, the authors obscured the issue of why the models failed. It may have been because the models neglected convection, or because they were one-dimensional. These one-dimensional models required vapor flux to condense at the bottom of each ice layer and sublime from the top. Essentially the snow was modeled as a series of horizontal ice plates and intervening air spaces, not realistic snow geometry. All the enhancement factors (with the exception of the work by Fedoseeva and Fedoseev [1988], who do not indicate the source nor method of measurement of their flux determinations) were determined by comparing the model results to only two measured values of mass transport. Thus, Trabant and Benson (1972) interpreted their mass transport measurements, which were nearly an order of magnitude greater than predicted by the models even with the enhancement factors, to indicate the presence of convection in the snow. However, the mismatch between the measured value of Trabant and Benson and values predicted by the models may have been the result of inherent problems with the one-dimensional models.

The failure of the one-dimensional diffusion models to predict the correct mass transport, or to predict it only with heuristic diffusion coefficients, led to the development of two- and three-dimensional models with more realistic snow geometry (Adams and Brown, 1983; Sommerfeld, 1983; Colbeck, 1983a; Gubler, 1985; Christon et al., 1987). But, unlike the older one-dimensional models which predicted mass transport between snow layers (i.e. layer-to-layer flux gradients), the multi-dimensional models were microscopic in scale. They predicted the local growth rate of snow grains. To simplify the mathematics of these multi-dimensional models, all of the authors assumed that the flux gradients between snow layers were zero, making the one- and multi-dimensional models contradictory. This assumption was justified by citing the experiments of Marbouty (1980) and the observations of Armstrong (1985), both of whom found that the mass of a snow layer in which there was active grain growth (and shrinkage) remained constant. The work presented in this study, and that of Trabant and Benson (1972), clearly indicates that there is mass flux from one layer to another.

The existing multi-dimensional models are untested. Some incorporate undetermined enhancement factors. The models require detailed geometric measurements of snow grain clusters which are not available, and they predict growth rates, of which there are few reliable measurements. As a result, the models are not yet useful for determining whether observations of mass transport or grain growth in snow can be explained solely by diffusion with realistic snow geometry, or if transport by convection must be included.

The fundamental problem is that none of the models can be adequately tested because there are so few measurements of mass transport or grain growth in snow. The measurements which do exist (Table 1-1) are difficult to compare because they were made using various experimental techniques on different types of snow subjected to different temperatures and temperature gradients. The measurements listed in Table 1-1 have been separated into a) flux gradients between

TABLE 1-1: Layer-to-layer mass flux gradients (a) and grain growth rates (b) for snow.

a) MEASURED VALUES OF LAYER-TO-LAYER FLUX GRADIENTS

SOURCE	EXPERIMENTAL METHOD	COMMENTS	FLUX GRADIENT [$kg\ m^{-2}\ s^{-1}\ m^{-1}$]	TEMP. [^{o}C]	T-GRAD. [$^{o}C\ m^{-1}$]
Kondrat'eva (1954)	Calculated from change in thermal diffusivity.	Indirect calc. subject to large errors.	190×10^{-6}	-1 to -10	-20
Yosida et al. (1955)	Mass change of snow in wire cages in snow cover.	Wire cages may have enhanced temp. gradients.	1×10^{-6}	0 to -6	-4
de Quervain (1958)	Mass change of block of snow in controlled environment.	Got similar values for tests with no temp. grad. as with strong grad.	10×10^{-6}	-5	0 to -70
Giddings & La-Chapelle (1962)	Observed change in density of natural snow layers.	Did not account for effects of snow compaction.	---------	-3 to -8	-30 to -50
Yen (1963)	Change in density of ventilated snow. Results extrapolated to 0 flow.	Data not presented in paper, but confirms Yosida et al.'s (1955) results.	1×10^{-6}	- 7 to -17	-70
Trabant & Benson (1972)	Comparison of observed change in density of natural snow to snow not subjected to temperature gradient.	Value may be too high due to differences in compaction rate of different snow types.	9×10^{-6}	VARIED	MAX -200
This study	Measured mass and thickness change of marked natural strata.	See Chapter 4 and Appendix V	3×10^{-6}	VARIED	MAX -200

b) MEASURED VALUES OF SNOW GRAIN GROWTH RATES

			GROWTH RATE [$kg\ s^{-1}$]		
de Quervain (1958)	Samples subjected to known gradients. Grain size est. by sieving and photos.	Grain size increases from 1 mm to 5 mm in 40 days.	1.6×10^{-11} (*)	-5	-70
Pahaut & Marbouty (1981)	Measured from photographs of disaggregated grains (?)	Grain size increases from 0.4 to 6 mm in 30 days.	4.6×10^{-11} (*)	-4	-40
Sommerfeld (1983)	Samples in plastic subjected to known grad. Grain size by stereology.	Grain size increases from 2 mm 3 mm in 28 days.	0.4×10^{-11} (*)	-4	-30
This study	Natural samples. Grain mass determined by sieving and stereology.	Grain size increases from 5 mm to 50 mm in 120 days.	0.8×10^{-11} (**)		VARIED Max -200

(*) growth rate calculated assuming spherical ice grains with diameter = grain size.
(**) growth rate calculated by methods indicated in Chapter 4 and Appendix VI.

snow layers and b) grain growth rates. The results span an order of magnitude for both types of measurements. Among the layer-to-layer flux determinations (Table 1-1a), one value (Giddings and LaChapelle, 1962) was based on measured changes in snow density, but failed to account for compaction. Another value (Yen, 1963) cannot be verified because the data have not been published. Still another value (de Quervain, 1958) is puzzling because the flux gradient was the same for blocks of snow both with and without imposed temperature gradients. The snow grain growth rates (Table 1-1b) were calculated from published values of changes in grain size by assuming spherical grains (except for this study, see Section 4.3). Accurate calculations would have required knowledge of grain shape as well as size, but these data were not available.

1.2.2 Diffusive Heat Transport in Snow

In the absence of convection, heat 1) is conducted through the network of ice grains, 2) is conducted through the air spaces in the pores, and 3) moves as latent heat when vapor diffuses from one grain to another. The effective thermal conductivity of the snow combines all three mechanisms. The thermal conductivity of air and ice are well known, and the effective conductivity of the snow can be measured, but the component of the effective thermal conductivity due to latent heat transport has yet to be calculated from the other components. The reason is that the complex geometry of pore spaces and ice grains produce unknown local temperature and vapor pressure gradients. Also, all three heat transfer mechanisms are poorly known functions of temperature. Only three studies have measured the effective thermal conductivity over a range of temperatures including low temperatures (≤-40°C) where vapor transport is negligible (Pitman and Zuckerman, 1967; Voitkovsky et al., 1975; this study, Section 3.2.2).

Several authors have attempted to evaluate the latent heat transported by the vapor flux using theoretical models. Yosida et al. (1955) predicted that about 40% of the total non-convective heat transport should be due to the vapor. Using the measured vapor transport, however, they found the obvious contradiction that more than 100% heat transport could be accounted for by the latent heat. Yen (1963) concluded that only 8% of the heat moved as latent heat, even at relatively high temperatures, whereas Woodside (1958), modeling ice spheres and air spaces, concluded that up to 48% of the heat was transferred with the vapor. De Quervain (1972) calculated latent heat transport as a function of temperature, snow density, and texture. He found that the percentage of the total heat transferred as latent heat increased with porosity, and that for low density depth hoar it reached values up to 45%.

1.2.3 Natural Thermal Convection in Snow

In addition to the studies of natural thermal convection in snow discussed in Section 1.1, Bories and Combarnous (1973) examined the theoretical likelihood of convection in snow on a sloping surface and concluded that there would be a general up-slope flow of air in the snow. Forced convection due to wind-pumping has also been shown to be possible on theoretical grounds (Clarke et al., 1987; Colbeck, in prep.). Data presented in Section 3.2.3 may confirm that wind-pumping occurs. Johnson et al. (1987), reporting results from the first year of this study, found natural thermal convection during a brief period of the winter of 1985. The prevalence of convection in other snow covers is not known.

1.3 General Characteristics of the Fairbanks Snow Cover

The Fairbanks snow cover is similar to that found in Siberia, Canada, and the rest of Interior Alaska, coinciding with the Taiga vegetation zone. These snow covers are colder, thinner, less dense, more permeable, of lower conductivity, and subject to more extreme temperature and vapor pressure gradients than snow covers in temperate latitudes and have been called Taiga snow (Pruitt, 1970, 1984; Benson, 1982). They evolve as winter progresses until they consist almost entirely of large depth hoar crystals (see Frontispiece).

1.3.1 The Climate of Fairbanks

Fairbanks, Alaska (64° 49′ N, 147° 52′ W, elev.:130 m) has a dry, continental, subarctic climate that is often windless (Searby and Branton, 1973). Below freezing temperatures usually last from October to April, and the seasonal snow cover can form any time after the beginning of October and persist until May (Figure 1-3) (see also Bryson and Hare, 1974). Because thaws occur rarely in winter, the snow cover is often free of ice lenses or melt features. The mean wind speed during the winter is less than 3 m s^{-1}.

1.3.2 Depth, Density, and Texture of the Fairbanks Snow Cover

The snow cover in Fairbanks reaches its maximum thickness by March; it averages 0.6 m with significant variation from year to year (Fig. 1-3 and Table 1-2). Typically, the snow cover is composed of a dozen or more individual snowfalls, many less than 0.1 m thick. Due to the windless conditions, most snowfalls are deposited with initial densities of less than 50 kg m^{-3}. These low density accumulations often consist of unbroken stellar snowflakes, sector plates, and spatial dendrites (Magono and Lee [1966] Types: P1e, P1f, R1d, R2b, P6a, P1c, P6c). The snowflakes quickly lose their points and become rounded grains as they settle and compact (Bader et al., 1939). Kinetic metamorphism (Colbeck, 1986) (also occasionally called TG metamorphism [Sommerfeld and LaChapelle, 1970]) quickly turns the grains into depth hoar which has an average density between 180 and 250 kg m^{-3} throughout the winter.

Observations of natural snow covers (Akitaya, 1974) and experimental work (Marbouty, 1980; Colbeck, 1983b) suggest that the temperature gradient must exceed 10 K m^{-1} to 25 K m^{-1} for depth hoar to develop. When temperature gradients are less than these

TABLE 1-2: Maximum snow depth and month of the maximum for Fairbanks, Alaska, 1950 to 1987. Note the extreme values in adjacent years 1970 and 1971.

DATE	DEPTH (m)	MONTH	DATE	DEPTH (m)	MONTH	DATE	DEPTH (m)	MONTH
1950	1.09	FEB.	1963	0.89	MAR.	1976	0.46	MAR.
1951	0.86	MAR.	1964	0.41	MAR.	1977	0.74	FEB.
1952	0.84	FEB.	1965	0.43	MAR.	1978	0.48	FEB.
1953	0.23	MAR.	1966	1.32	FEB.	1979	0.64	FEB.
1954	0.48	FEB.	1967	0.94	APR.	1980	0.41	FEB.
1955	0.56	MAR.	1968	0.84	JAN.	1981	0.53	FEB.
1956	0.97	MAR.	1969	0.74	MAR.	1982	0.46	APR.
1957	0.94	JAN.	1970	0.30	FEB.	1983	0.61	MAR.
1958	0.36	FEB.	1971	1.09	MAR.	1984	0.65	FEB.
1959	0.91	FEB.	1972	0.74	FEB.	1985	0.88	JAN.
1960	0.91	JAN.	1973	0.56	MAR.	1986	0.40	APR.
1961	0.38	JAN.	1974	0.56	MAR.	1987	0.42	JAN.
1962	0.94	FEB.	1975	0.79	JAN.			

Based on data supplied by the Arctic Environmental Information and Data Center, University of Alaska these values, snow crystals become rounded, approaching an equilibrium shape that minimizes the surface free energy (Frank, 1982; Sommerfeld and LaChapelle, 1970; Colbeck, 1982a, 1982b). Above the critical gradient, crystal growth dynamics, rather than vapor supply, apparently control the crystal growth process, and ornate depth hoar crystals develop with striated and stepped faces (Frank, 1982).

Temperature gradients in the Fairbanks snow cover often exceed the critical gradient by an order of magnitude or more for up to 150 days (Fig. 1-1). Temperature gradients can be as high as 500 K m^{-1}, particularly in early season when winter ground temperatures are the highest and the snow cover is the thinnest. At that time, the formation of depth hoar can be extremely rapid.

The extreme temperature gradients are the result of the low winter air temperatures and the relatively high snow-soil interface temperatures. The interface temperatures remain high because latent heat is released as moisture in the soil freezes. Interface temperatures rarely drop below -8°C, even when air temperatures are as low as -40°C.

Observations by Akitaya (1974) suggested that depth hoar does not form, regardless of the temperature gradient, if the initial snow density exceeds 350 kg m^{-3}. Because the snow depth in Fairbanks is usually less than a meter (Table 1-2), compaction of snow layers at the base of the snow cover resulting from the weight of the overlying snow is minimal and this critical density is rarely exceeded.

1.4 Organization of Chapters

This was primarily an observational study. The extensive experimental methods used in the study are described in Chapter 2. Observations of density, grain size, and temperature, used to calculate the mass and heat transport, are presented in Chapter 3. From the density observations it was possible to determine the layer-to-layer mass flux gradients; from the grain size data it was possible to estimate grain growth rates. These are discussed in Chapter 4, and then used in Chapter 5, where the evidence for convection is presented, and its importance is discussed. Chapter 6 is a summary of results and recommendations for future work.

2 METHODS

This study was conducted in a natural (as opposed to a laboratory) setting because its purpose was to determine if convection occurred in a natural snow cover. Severe arctic weather and other natural events (e.g., a moose treading perilously close to the instrumented snow site) had to be dealt with, and measurement techniques with minimal thermal and structural impact on the snow had to be devised.

Convection was detected primarily by its perturbation of the temperature field in the snow, and secondarily by changes in heat flow at the snow-soil interface. Temperature and heat flow were measured more than 3×10^6 times over three winters using thermistor arrays in the snow and soil, and heat flow meters at the snow-soil interface. Thermistors and heat flow meters were installed at the end of summer and allowed to be buried by the winter snowfall, thus assuring that the snow in which the instruments were buried was representative of the general snow cover.

Mass transport and snow metamorphism were studied by making measurements of density, stratigraphy, and grain size. Fifty-five snow pits were dug over three years, and over 600 density measurements were made. Samples from each snow pit were sieved and photographed to determine grain size. Other measurements, such as soil moisture profiles and thermal conductivity of the snow, were made to complete the description of the snow and its environment.

Julian dates were used during the course of the study and appear on several graphs. Table 2-1 is a Julian-to-conventional date conversion chart.

2.1 Experimental Setting

Observations were made at the University of Alaska Agricultural Experiment Station, the same site used by Trabant and Benson (1972) for their work. The site is a level, cultivated field with silty soil.

2.2 Measurements of the Physical Characteristics of the Snow

2.2.1 Density, Stratigraphy, Depth, and Compaction Measurements

Snowfall accumulated on six white tables (1 m-by-2 m) 1.5 m above the ground, and on a 20 m-by-20 m area of ground surface. The ground was leveled, raked and rolled prior to the first snowfall. Once a month throughout the winter snow pits were dug in the snow on the ground and the tables. This setup, which was similar to that used by Kojima (1959), Trabant and Benson (1972), and Armstrong (1985), allowed a comparison to be made between the snow on the ground, which was subjected to strong temperature gradients, and the snow on the tables which was subjected to only weak temperature gradients which had no preferred direction.

Density, snow depth, and stratigraphy were measured in each snow pit using procedures described in NRC Canada Memorandum No. 31 (1954),

TABLE 2-1: A Julian date vs. conventional (Calendar) date conversion chart. For leap years, add 1 after February 28.

JULIAN DATE CALENDAR

(PERPETUAL)

Day	Jan	Feb	Mar	Apr	May	June	July	Aug	Sep	Oct	Nov	Dec	Day
1	001	032	060	091	121	152	182	213	244	274	305	335	1
2	002	033	061	092	122	153	183	214	245	275	306	336	2
3	003	034	062	093	123	154	184	215	246	276	307	337	3
4	004	035	063	094	124	155	185	216	247	277	308	338	4
5	005	036	064	095	125	156	186	217	248	278	309	339	5
6	006	037	065	096	126	157	187	218	249	279	310	340	6
7	007	038	066	097	127	158	188	219	250	280	311	341	7
8	008	039	067	098	128	159	189	220	251	281	312	342	8
9	009	040	068	099	129	160	190	221	252	282	313	343	9
10	010	041	069	100	130	161	191	222	253	283	314	344	10
11	011	042	070	101	131	162	192	223	254	284	315	345	11
12	012	043	071	102	132	163	193	224	255	285	316	346	12
13	013	044	072	103	133	164	194	225	256	286	317	347	13
14	014	045	073	104	134	165	195	226	257	287	318	348	14
15	015	046	074	105	135	166	196	227	258	288	319	349	15
16	016	047	075	106	136	167	197	228	259	289	320	350	16
17	017	048	076	107	137	168	198	229	260	290	321	351	17
18	018	049	077	108	138	169	199	230	261	291	322	352	18
19	019	050	078	109	139	170	200	231	262	292	323	353	19
20	020	051	079	110	140	171	201	232	263	293	324	354	20
21	021	052	080	111	141	172	202	233	264	294	325	355	21
22	022	053	081	112	142	173	203	234	265	295	326	356	22
23	023	054	082	113	143	174	204	235	266	296	327	357	23
24	024	055	083	114	144	175	205	236	267	297	328	358	24
25	025	056	084	115	145	176	206	237	268	298	329	359	25
26	026	057	085	116	146	177	207	238	269	299	330	360	26
27	027	058	086	117	147	178	208	239	270	300	331	361	27
28	028	059	087	118	148	179	209	240	271	301	332	362	28
29	029		088	119	149	180	210	241	272	302	333	363	29
30	030		089	120	150	181	211	242	273	303	334	364	30
31	031		090		151		212	243		304		365	31

SIPRE Instruction Manual No. 1 (1962), and by Benson (1962). Metal tubes (0.5 liter volume) were inserted horizontally into the wall of a snow pit from which a core of snow was removed and weighed. The tubes, which were 60 mm in diameter, often sampled across several snow layers. This was undesirable because the density of individual layers often differed. Therefore, in 1986 and 1987 a 30 mm high, 0.1 liter box sampler, similar to the Hydro-Tech/Institute of Low Temperature Science sampler (P. Taylor, personal communication), was used. Despite its smaller volume, the box sampler measured density to the same accuracy as the larger tubes (Carroll, 1977), and the problem of sampling several snow layers at once was avoided.

It became apparent in 1986 that measuring the snow density more frequently than once a month might improve the resolution of the mass transport calculations. Single density measurements were reproducible to ± 5% for a density of 200 kg m^{-3}. This was comparable to the reproducibility determined by Benson (1962) for Greenland snow. Unfortunately, this reproducibility indicated that the measurements were not accurate enough to differentiate the small changes in density that would occur over intervals of time shorter than a month. To improve the accuracy, multiple snow pits were measured at each site and the average density determined from several measurements for the same layer. This required processing more samples, so the efficiency of the process was improved by using standardized tare weight containers and an electronic balance. The accuracy of the average density measurements was a function of the snow texture. In older snow which had not been subjected to strong temperature gradients (i.e., snow on tables), and in fine-grained new snow, average density measurements were estimated to be accurate to ± 1% for a density of 200 kg m^{-3}. For depth hoar, which was difficult to sample, the accuracy was estimated to be ± 2.5%. Lateral density variations, particularly in depth hoar, probably produced further uncertainty on the order of ± 2.5%, giving a total accuracy of about ± 4% for depth hoar.

The metamorphism of the original snow into depth hoar obliterated all primary natural stratigraphic markers by mid-winter, so it was necessary to introduce artificial markers into the snow in order to monitor the compaction of snow layers. In 1986 this was done by laying colored yarn on the snow surface after major snowfalls. This was unsuccessful. In 1987, powdered paint (Binney & Smith Artista II tempera paint) was spread on the snow surface after major snowfalls with a flour sifter, producing a layered effect. The paint was inert and could be spread in thin layers that had little or no effect on the snow permeability; it was unlikely to have had any effect on vapor flow through the snow.

Snow compaction was computed from measured changes in thickness of the layers defined by the paint horizons. These could be measured accurately to ± 3 mm. In 1987, cumulative compaction of snow layers was also measured hourly using a data logger and the snow settlement device shown in Figure 2-1 (see Section 3.1.1).

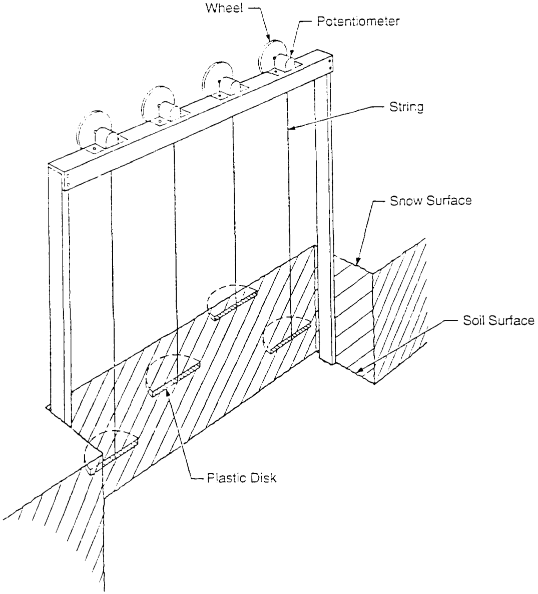

FIGURE 2-1: An apparatus for measuring the snow settlement. Four plastic disks were placed on the snow surface after significant snow falls. These were connected by strings to potentiometers turned by pulley wheels. The resistance of each potentiometer was calibrated with the height of the disk and read hourly by the data logger.

2.2.2 Grain Descriptions and Photography

Snow crystals and grains were examined in-situ using a hand lens. The entire snow cover was often photographed in "thin section" by cutting two adjacent pits and shaving the intervening wall until it was translucent (Benson, 1962). Under these conditions, the texture of the snow was highlighted. Grains from disaggregated snow samples were photographed in a cold room using an Olympus OM-4 35 mm camera mounted on an American Optical stereo-microscope. The best results were achieved by photographing grains on a glass plate illuminated by transmitted light (see Section 3.1.3).

The photography had two purposes: to determine grain growth rates and to document changes in crystal type during depth hoar development. A limited amount of photography was done using polarized light to verify crystal axes orientation, which could usually be determined from both hand lens examination and regular photographs due to the euhedral morphology of most crystals.

2.2.3 Grain Size Measurements

Sieving and photography were used to determine grain growth rates. Sieving is an old method of measuring grain size that seems to have gone out of favor because it is difficult to determine how much damage is done to the snow grains during the sieving. Several authors (Bader et al., 1939; Benson, 1962; Keeler, 1969; Fukue, 1977; Granberg and Wener, 1986) have analyzed sieved snow and most acknowledge that sieving can disaggregate snow grains and break single crystals, potentially biasing the results towards smaller sizes. In this way, sieving snow differs from normal sieve analysis of rock sediments, but as in sieving other sediments, operator error (Folk, 1955) and errors due to the sieving process itself (Griffiths, 1953; Mizutani, 1963; Ludwick and Henderson, 1968) also affect the results. Stereological measurements from photographs (Kry, 1975; Gubler, 1978) have been in use more recently, but they are hard to interpret, and are time consuming and difficult to make on snow composed of large, ornate depth hoar grains.

Despite its shortcomings, sieving is fast and easy to perform in the cold, and, if applied in a consistent manner, is adequate to measure the changes in grain mass necessary to determine grain growth rates. It was used in this study for these reasons. In standard sieve analysis, a particle size distribution is determined from the weight fraction of a sample found in each sieve (Royse, 1970; Blatt et al., 1972; Friedman and Sanders, 1983). An underlying premise in the standard method is that the size of a grain is related to the size of the sieve mesh opening. For spheres and other simple, regular shapes this is true (Ludwick and Henderson, 1968). For highly convoluted shapes like snow grains, no such relationship exists, so the sieving cannot be used to measure snow grain size accurately. However, by sieving a representative sample of depth hoar, then weighing each grain in each sieve, a direct relationship between sieve mesh opening and the average mass of a grain which remained in that sieve was developed (see Section 3.1.2). It was then possible to use sieving to determine the grain mass distribution for a snow layer at several times during the winter. The distributions could be used to determine the number of grains in a snow sample (see Section 3.1.2 and Appendix IV).

Samples for sieving were taken with a 0.5 liter tube from three fixed heights in the snow cover on both the ground and the tables each time snow pits were dug during the winter of 1987. Each sample was transferred to the top sieve in a stack of 9 sieves, which were at ambient outdoor temperature (usually < $-20^{\circ}C$). The sample was gently disaggregated by moving it across the sieve mesh with a gloved hand. The sieve stack was then covered and agitated by hand for 30 seconds. The weight fraction in each sieve was weighed using an electronic balance.

Stereological measurements of grain size were made from photographs of disaggregated grains. For each snow layer, two or more photographs showing a grain population of between 50 and 200 grains were digitized (by tracing the grain outlines with a stylus) using a ZeissPlot stereological system. Descriptive statistics for each population (corresponding to a particular snow layer at a particular time) were calculated by the Zeiss software. The results included mean grain size and shape parameters. Sufficient stereological measurements were made in order to compare sieving and stereological methods of measuring grain size (see Section 3.1.2 and Fig. 3-7).

2.3 Heat Flow and Temperature Measurements

2.3.1 Heat Flow at the Snow-Soil Interface

Heat flow meters (HFMs) (Thermonetics Model H11-18-3-sfh glass phenolic thermopiles) were embedded in the soil surface prior to the first snowfall. Voltage from the HFMs was recorded hourly, and the output was converted to heat flow units using a temperature-compensating factory calibration. The mismatch in the thermal conductivity between the HFMs and the soil produced a "focusing" of heat flow through the meters which was a function of the aspect ratio of the HFMs and the ratio of the thermal conductivity of the HFMs (k_m) to the thermal conductivity of the surroundings (k_s). The correction for this mismatch was made using Schwerdtfeger's (1970) equation:

$$(2\text{-}1) \qquad Q_s = Q_m \left(1 + H \left[\frac{k_m/k_s - 1}{k_m/\ k_s}\right]\right)$$

where H is a geometric characteristic of the HFMs determined empirically for different aspect ratios, Q_s is the heat flux at the snow-soil interface, and Q_m is the heat flux measured by the meter. For the silt at the experimental site, k_s probably varied between 2.0 and 0.5 W m^{-1} K^{-1} as the winter progressed and the soil moisture froze (Johnston, 1981); k_m was 0.17 W $m^{-1}K^{-1}$. Thus, Q_s was between 2.9 and 1.3 times Q_m. Because k_s was not known precisely, it was decided that an average multiplication factor of 2 should be used.

The corrected readings of the HFMs are reported in this study as the heat flow at the snow-soil interface, but it is not clear that this is exactly what the HFMs measured. If the system had been purely conductive with no vapor transport, the measured value would be the vertical heat flow at the interface. However, with air convecting in the snow, and a continuous vapor flux out of the soil, the heat flow should have been a function of location with respect to the air circulation pattern and the soil moisture content. The component of the heat flow due to the the release of latent heat resulting from condensation of vapor on the bottom of the impermeable meters was relatively small and can be ignored. Condensation of about 5 mm of ice beneath the HFMs by the end of each winter released a latent heat flow of less than 1 W m^{-2}, while the measured heat flow varied between 5 and 30 W m^{-2} (see Section 3.2.1). The close agreement in the readings between two (1985) and three (1986-1987) HFMs spaced several meters apart (see Fig. 3-16) suggests that the heat flow may have been relatively constant along the snow-soil interface, despite convection. Therefore the heat flow values measured by the HFMs are useful, even if they are not exactly equal to the vertical heat flow at the snow-soil interface.

2.3.2 Temperature Measurements in the Snow and Air

Temperature measurements were made by strings of thermistors which were suspended before the first snowfall of the winter and were allowed to be buried by snow. The number and arrangement of the thermistor strings was improved each year in an effort to optimize coverage of the temperature field in the snow. The thermistor strings were configured as follows:

1985 (Fig. 2-2): Two horizontal and one vertical thermistor string with a total of 103 thermistors were set up as described by Johnson et al. (1987).

1986 and 1987 (Fig. 2-3): Six horizontal and five vertical thermistor strings were set up to define a horizontal plane 1.0 m wide by 2.5 m long and a vertical plane 0.8 m wide by 0.7 m high. Each horizontal string incorporated 26 thermistors. They were installed 0.2 m above the ground in 1986, and 0.15 m in 1987; the smaller height in 1987 was to ensure that the thermistors would be covered earlier in the winter. The vertical strings actually consisted of 12 horizontal strings installed one above the other to create a vertical plane. In 1987, five thermistors were installed at the soil surface in order to measure the temperature of the snow-soil interface (Fig. 2-3). These were partially embedded in the soil, so that when covered by snow they would be half in the snow and half in the soil.

Once the thermistor strings were covered by snow, they were cut free of their supports, so that they could settle with the snow cover. This was preferable to having them settle in catenary curves, which would have resulted had they not been cut free. An exception was the vertical thermistor string used in 1985, which remained fixed to its supports. At the end of the season, the thermistors were excavated from the snow, and their positions carefully measured with respect to the ground, snow surface, and neighboring thermistors. In 1987,

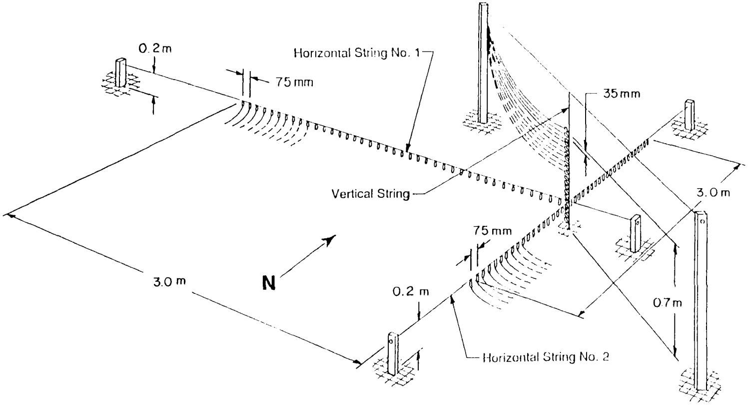

FIGURE 2-2: The thermistor array used in 1985. It consisted of two horizontal and one vertical string. Horizontal strings incorporated 41 thermistors each, and the vertical string inccrporated 21 thermistors.

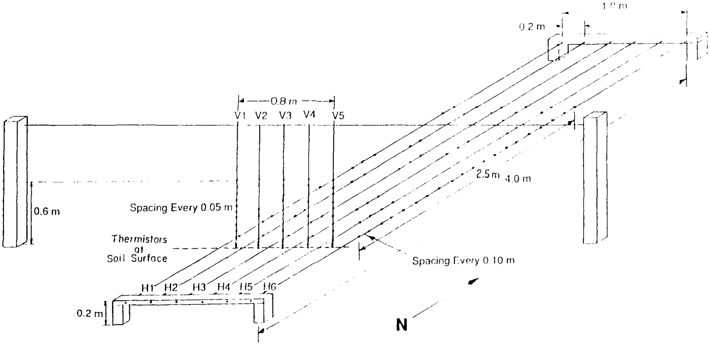

FIGURE 2-3: The thermistor array used in 1986 and 1987. It consisted of 6 horizontal strings, each incorporating 26 thermistors, and 5 vertical strings (actually made from horizontal strings) incorporating 12 thermistors each. In 1986, the horizontal strings were set 0.20 m above the ground. In 1987, they were set 0.15 m above the ground.

thermistor heights could be determined to ± 3 mm throughout the winter by using the snow settlement data (see Section 2.2.1), but insufficient settlement measurements were made in 1986, so the vertical positions of the thermistors were not known with enough accuracy to allow extensive analysis.

There was no discernible effect on the snow cover, or processes occurring within it, due to the presence of thermistor strings. The thermistors were small, and the thinnest possible leads were used so that minimal snow was intercepted during snowfalls, ensuring a natural snow cover surrounding the thermistors. Leads were installed horizontally to eliminate vertical heat flow along wires. When excavated, lead wires were found to have remained nearly horizontal despite settling of the snow. During the excavation, particular attention was paid to the nature of the contact between the thermistor beads and the snow around them. Thermistors were always in contact with the snow and often encased by ice crystals.

The thermistors were small glass beads 10 mm long and 3 mm in diameter (Veco Engineering Model T32A11/21). Each thermistor was factory calibrated at 25°, 0°, and -40°C. From this calibration, the Steinhart-Hart equation (YSI, 1984) was used to convert resistance to temperature with an accuracy of ± 0.01°C. At the end of each winter, the ice point calibration of each thermistor was checked in an ice bath. Ice point calibrations generally agreed with the factory calibration within 3 to 5 ohms out of 6000 ohms, indicating that loss of calibration through time was less than 0.1%, corresponding with an uncertainty of ± 0.02°C at 0°C and less for lower temperatures.

Instrument error and Joule heating (Appendix II) resulted in a further uncertainty of ± 0.02°C. A conservative estimate of the total uncertainty is ± 0.03°C.

The data collected in 1985 indicated that it was critical for the ground surface below the horizontal strings to be smooth and level in order to minimize perturbations in the temperature field due to variations in the distance between the thermistors and the ground. In 1985, the substrate was the bare soil, raked and rolled smooth, leaving undulations with amplitudes of 20 to 40 mm and wavelengths of about 0.5 m on the soil surface. In 1986 the upper 100 mm of soil was tilled and then raked and hoed. The surface was wetted and allowed to settle, then small depressions were filled, resulting in a surface with undulations less than 20 mm high. In 1987, a leveling frame consisting of boards dug into the ground and leveled to 1 mm was used to smooth and level a 20 mm thick layer of sand spread over the silt. This was accomplished by running a stiff metal rail across the leveling boards and the surface of the sand, and resulted in an undulating surface with an amplitude less than 4 mm.

The number of thermistors (between 103 and 214) and the need for a continuous temperature record throughout the winter required an automatic data logging system. The system consisted of a computer (Hewlett Packard [HP] 86C) controlling two data loggers (HP 3421A) and one multiplexer (HP 3497A). The system was contained in an insulated box in an unheated building. Waste heat from the equipment kept the mean temperature in the box at 20° ± 5°C. Data were stored on 3 1/2" floppy discs, each of which could hold a 10-day record. The entire system was connected to a battery power supply that maintained power during frequent power outages. Transducers were logged every 15 minutes in 1985 and every hour in 1986 and 1987, producing 16 Mbytes of data.

The raw data collected by the data logging system were transferred via disc-to-disc transfer to an HP 200 series computer for reduction. A data handling and graphics program written by S. Perkins of USA-CRREL (Perkins, 1987), which could output the temperature and heat flow data in graphic form, was used for initial data handling.

Air temperature was measured by a thermistor housed in a standard meteorological shelter. Wind speed and direction were measured 200 m from the experimental site by a standard anemometer and wind vane. The wind measurements were part of a separate experiment run by J. J. Kelley and J. H. McBeath, who generously shared their data.

2.3.3 Temperature Measurements in the Soil

Ground temperatures were measured to a depth of approximately 2 m throughout the study using a string of Fenwal thermistors. The thermistor string was attached to a post installed in an augered hole which was then back-filled and compacted. For protection, the thermistors were encased in heavy shrink tubing. The Fenwal thermistors (Model UUA-33J1) were not factory calibrated. A single ice point calibration and the factory-supplied standard curve for resistance vs. temperature were used, giving an accuracy of $\pm$ 0.2^{o}C.

2.3.4 Measurements of the Thermal Conductivity of the Snow

The thermal conductivity of the snow was measured using a needle probe as described by Jaafar and Picot (1970), Lange (1985), and Sturm and Johnson (1987) (Appendix III). Briefly, a 0.1 m diameter by 0.3 m long copper cylinder was used to take a horizontal snow sample. The sample was transported to a laboratory where it was placed in a jacket around which cooling fluid circulated. The sample was brought to the desired temperature and stabilized. A thin steel needle, 0.2 m long and 1.5 mm in diameter, containing a helical resistance heating wire and a thermocouple, was inserted along the axis of the sample. The needle was heated for 5 to 10 minutes while its temperature was monitored. The heater was then shut off and the cooling of the sample was monitored. From the heating and cooling curves, two independent values of the effective thermal conductivity could be determined (Blackwell, 1954; Lachenbruch, 1957; Jaeger, 1958; von Herzen and Maxwell, 1959; Pratt, 1969; McGaw, 1984; Lange, 1985).

Three measurements of thermal conductivity were made in 1985 and 14 in 1986. In 1987, samples from heights of 0.07, 0.10 and 0.26 m were measured five times during the winter to determine changes in thermal conductivity resulting from snow metamorphism (see Section 3.2.2). Measurements were made over a range of temperatures in order to assess temperature dependence. One measurement was made on a sample cooled in liquid nitrogen (-196°C) to determine the thermal conductivity in the absence of latent heat transport due to vapor flux.

2.4 Miscellaneous Measurements

2.4.1 Soil Moisture Content

In 1985 and 1986, near-surface soil moisture was monitored by drying samples taken from the top 50 mm of the soil. In 1987, soil moisture was monitored to a greater depth. The soil was cored to a depth of 1 m; the core was sectioned and weighed, then dried to determine the moisture content. This was done three times during the winter.

2.4.2 Vapor Flux Out of the Soil

An impermeable sheet (polyethylene plastic, 5 m-by-5 m) was laid on the soil surface just after the last rain of autumn and at the start of freezeup. Moisture trapped under the sheet was used to determine the moisture flux from the soil to the snow. Periodically through the winter, a 0.3 m-by-0.3 m square of the sheet was cut away, and the ice under the sheet collected and weighed. The upward flux of vapor from the soil could be determined from this weight (see Section 3.1.1).

2.4.3 Air Permeability

Air permeability of the snow was measured several times in 1985 and 1986 using an air permeameter similar to the one described by Bender (1957). In 1987, E. Chacho and J. Johnson of USA-CRREL made a series of measurements using a newly designed air permeameter which used solid-state pressure transducers, low laminar air flow, and a double-walled sample holder to minimize edge effects. Preliminary results were reported by Chacho and Johnson (1987) and are summarized in Appendix I. Air permeability was measured at the same time as thermal conductivity.

3 OBSERVATIONS

The observations fall into two categories: 1) observations of the physical characteristics of the snow and 2) measurements related to heat transport in the snow. The first category includes measurements of snow density, stratigraphy, grain size, and observations of the snow texture. These were made to calculate the mass transport in snow, to determine snow grain growth rates, and to document the snow metamorphism and relate it to convection, if possible (Chapter 4). The second category includes measurements of heat flow, thermal conductivity, and snow temperature. Temperature measurements made using the horizontal thermistor strings provided the primary evidence from which the presence of convection was inferred (Chapter 5). The curvature of the vertical temperature profiles, measured using the vertical thermistor strings, proved useful for differentiating between two types of convection. Heat flow, usually a good indictor of convection, proved difficult to measure in a meaningful manner (see Section 2.3.1), but the measurements are included for completeness.

3.1 Observations of the Physical Characteristics of the Snow

3.1.1 Snow Density

During every winter for which there are data, snow on the ground in Fairbanks developed a density profile in which the density decreased (or was constant) with depth. The density increased with depth in snow on the tables, which was not subjected to strong temperature gradients and did not metamorphose into depth hoar (Figs. 3-1, 3-2). Comparable snow layers on the ground and table rarely had the same density. In particular, the basal snow layers on the ground were always less dense than the basal layers on the tables, and the top layers of snow on the ground were denser than the comparable layers on the tables (Fig. 3-2). The density contrasts increased through the winter, reaching a maximum in March or April.

The striking difference between the two density profiles was interpreted by Trabant and Benson (1972) to be the result of a net transfer of mass from the base to the top of the ground snow cover by an upward-directed vapor flux. The present investigation supports this interpretation (see Section 4.2.1).

The upward-directed vapor flux could be measured directly only at the snow-soil interface. There the flux, measured by trapping and weighing the moisture which accumulated under an impermeable sheet (see Section 2.4.2), averaged 2.6×10^{-7} kg $m^{-2}s^{-1}$ (Table 3-1). This average was confirmed by measuring the change in the water content of the soil under snow where there was no impermeable sheet. During each of four winters, little or no change in the moisture content of the top meter of the soil was observed, with the exception of a surface layer, 20 to 30 mm thick, which became desiccated and dusty. This layer dried from 40% to less than 10% moisture by dry weight, releasing an average upward-directed vapor flux of 3×10^{-7} kg $m^{-2}s^{-1}$.

Individual snow layers on the ground and on the tables densified rapidly during the initial three to four weeks after they were

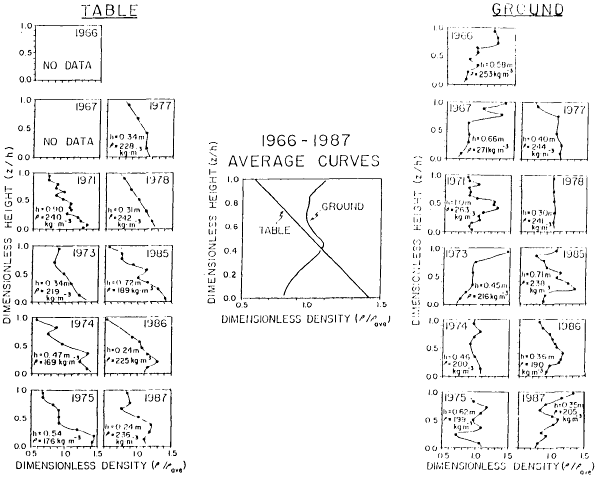

FIGURE 3-1: End-of-winter density profiles for the Fairbanks snow cover, 1966-1987. Snow on tables not subjected to strong temperature gradients did not metamorphose into depth hoar as did snow on the ground. Average density curves show distinct differences between ground and table profiles. The total snow depth (h) and average density (ρ) is listed for each snow pit.

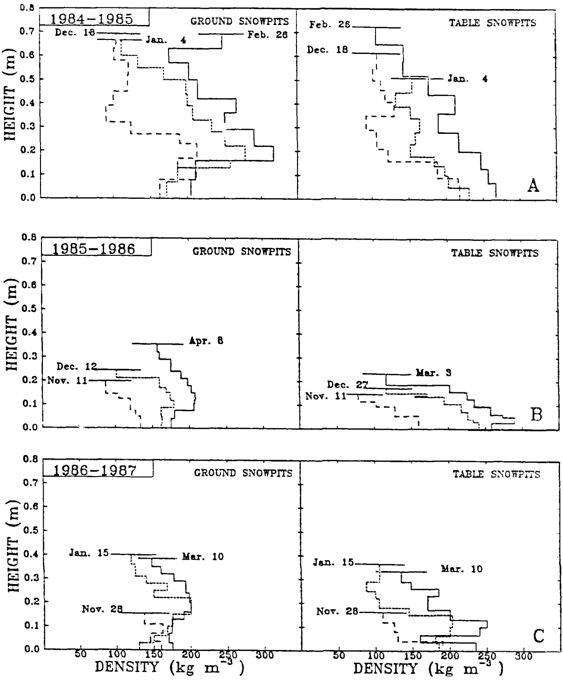

FIGURE 3-2: Fairbanks snow density profiles, 1985-1987. For each year, an early-, mid- and late-winter density profile is shown for the snow cover on the ground and the tables.

TABLE 3-1: Upward-directed vapor flux at snow-soil interface. Values were obtained by collecting the ice which formed under an impermeable sheet.

Year	Flux [kg m^{-2} s^{-1}] X 10^{-7}
1967-68*	3.2
1968-69*	2.8
1969-70*	3.5
1973-74**	1.7
1985-86	1.8
1986-87	2.6

*Trabant and Benson (1972)
**Benson, unpublished

deposited. This was followed by a second, longer period during which their rate of densification decreased (Fig. 3-3). In 1987, 10 snow layers on the ground (Table 3-2) and the equivalent 10 layers on the tables were marked with powdered paint (see Section 2.2.1), and the density and thickness of the layers were measured 13 times during the winter. For all layers, snow on the ground densified at a slower rate than the equivalent layer of snow on the tables during both the initial period of rapid densification and the secondary period of slower densification. Several snow layers on the ground experienced little or no change in density after the initial period was over. In these cases, the density of the snow layers remained constant for nearly 80 days (Fig. 3-3).

The snow layers compacted and settled as they densified. The compaction (change in thickness) of individual snow layers was determined by measuring the distance between paint layers (see Section 2.2.1). The layers changed thickness at a rate of 0.5 to 1.0 mm day^{-1} at first, but the rate of change decreased to less than 0.1 mm day^{-1} several weeks after the snow was deposited (Fig. 3-4). For several snow layers on the ground, layer thickness remained constant (within the measurement error) following the initial period of rapid compaction (Table 3-3). Equivalent layers of snow on the tables compacted twice as fast as snow on the ground during the initial period of rapid compaction (Fig. 3-4). The difference in compaction rates was the result of differences in the texture of the snow on the tables and on the ground. The snow on the tables was composed of small (about 1 mm), rounded grains which were well-bonded,

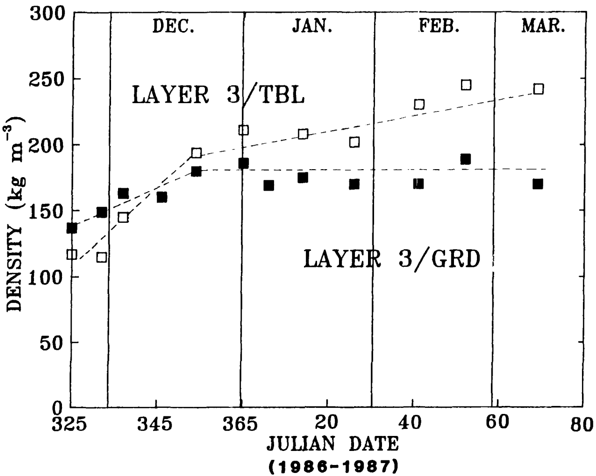

FIGURE 3-3: Densification of snow on the ground (GRD) and on the tables (TBL). Snow layer 3, which was at a height of 0.10 m, is shown because it was typical of the 10 layers which were measured. Both table and ground layers densified rapidly at first, followed by a period of less rapid densification.

TABLE 3-2: Density of 10 snow layers, 1987. Layer heights are given in Table 3-3.

DATE	21NOV	28NOV	3DEC	12DEC	20DEC	31DEC	6JAN	14JAN	26JAN	10FEB	21FEB	10MAR.
ELAPSED DAYS	0	7	12	21	29	40	46	54	66	81	115	145
LAYER 10									119	116	115	145
LAYER 9								125	170	168	175	164
LAYER 8							072	150	178	171	196	191
LAYER 7						120	111	151	161	192	200	183
LAYER 6					200	195	185	191	200	196	185	198
LAYER 5					194	187	182	200	201	202	218	192
LAYER 4			110	167	191	190	191	198	191	200	200	179
LAYER 3	120	135	163	160	180	186	169	175	170	170	189	170
LAYER 2	158	145	161	155	168	160	170	155	175	168	168	162
LAYER 1	172	155	183	160	151	173	162	145	168	172	157	171

all densities: kg m^{-3}

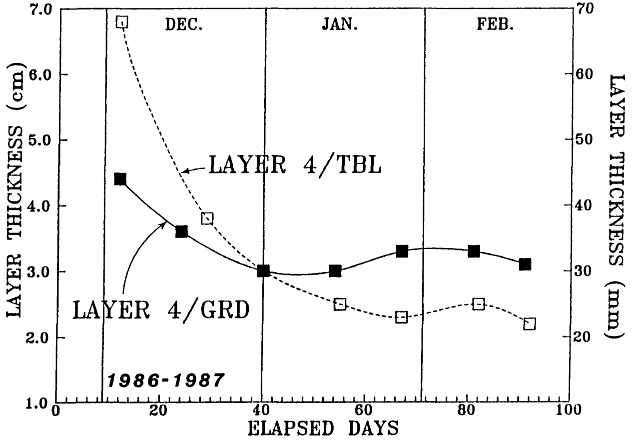

FIGURE 3-4: Compaction of snow on the ground (GRD) and on the tables (TBL). Layer 4, which was at a height of 0.14 m, was chosen because it was typical of the 10 layers which were measured in 1987. Initial thickness of the snow layer on the table was greater than the snow layer on the ground, but the layer on the table compacted more rapidly due to differences in snow texture (see text). Time is measured from the day the snow was deposited.

TABLE 3-3: Snow layer thicknesses and heights, 1987.

DATE	21NOV	28NOV	3DEC	15DEC	31DEC	14JAN	27JAN	10FEB	21FEB	11MAR.
ELAPSED DAYS	0	7	12	24	40	54	67	81	91	110
							36.5	36.3	34.0	34.0
LAYER 10							2.1	1.1	0.9	1.0
						37.8	34.4	35.2	33.1	33.0
LAYER 9						5.3	5.4	4.5	4.6	5.2
						32.5	29.0	30.7	28.3	27.8
LAYER 8						8.5	6.8	7.3	6.1	6.1
					25.8	24.0	22.2	23.4	22.2	21.7
LAYER 7					4.8	3.5	2.4	3.2	2.5	2.4
				22.0	21.0	20.5	19.8	20.2	19.7	19.3
LAYER 6				2.0	2.5	2.3	2.3	2.0	2.2	1.5
				20.0	18.5	18.2	17.5	18.2	17.5	17.8
LAYER 5				3.2	2.5	2.2	2.1	2.3	2.1	2.8
			20.0	16.8	16.0	16.0	15.4	15.9	15.4	15.0
LAYER 4			4.4	3.6	3.0	3.	3.3	3.5	3.1	2.7
	15.0	15.5	15.6	13.2	13.0	13.0	12.1	12.4	12.5	12.3
LAYER 3	6.7	7.5	6.8	5.0	5.0	5.5	4.3	4.9	4.6	5.3
	8.3	8.0	8.8	8.2	8	7.5	7.8	7.5	7.4	7.0
LAYER 2	2.3	2.0	1.8	2.3	2.0	--	--	1.5	2.1	2.4
	6.0	6.0	7.0	5.9	6.0	?	?	6.0	5.8	4.6
LAYER 1	6.0	6.0	7.0?	5.9	6.0	?	?	6.0	5.8	4.6
	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0	0.0

*********** KEY **********

TOP OF LAYER: —18.5 cm—
LAYER THICKNESS: 2.5 cm
BOTTOM OF LAYER: —16.0 cm—

compared to the large (5 to 20 mm), ornate, poorly-bonded depth hoar grains found on the ground. Kojima (1959, 1966), Bergen (1978), and Armstrong (1985) have reported that depth hoar, although brittle and weak in shear, is more resistant to compaction than snow of other textures with the same density.

Snow settlement was measured automatically by the movement of plastic disks embedded in the snow (see Section 2.2.1 and Fig. 2-1). The settling of the disks, the paint layers, and measurements of the total snow depth, compiled on Figure 3-5, agree closely.

3.1.2 Grain Size

Snow grain size is difficult to measure and varies with the method of measurement because 1) there is no standard definition of a snow grain, and 2) snow grains have irregular shapes. The second point is particularly true for depth hoar grains. Bader et al.(1939) and Sommerfeld and LaChapelle (1970) have proposed definitions of snow grains. Using the latter authors' definition, snow grains are considered to be the fundamental unit of the snow cover. They are either single ice crystals or aggregates of well-bonded ice crystals separated from the rest of the snow by weaker and fewer bonds than in the aggregate itself. Thus, sieving or disaggregation for photography generally breaks the snow into grains as opposed to breaking the grains apart. This propensity to break into grains is particularly prevalent in depth hoar because of the fragile nature of the connections between grains. Because the Fairbanks snow cover was

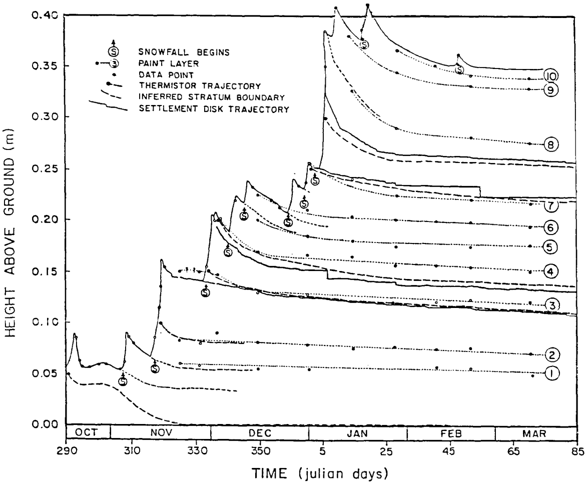

FIGURE 3-5: A stratigraphic record of the snow cover of 1987. The solid lines were measured by plastic disks (see Fig. 2-1) attached to potentiometers which sometimes "stuck" then released, giving the record characteristic steps. The settling trajectories of thermistors (bold, dashed lines) have been interpolated from the trajectories of the disks. Dotted lines show the settling trajectories of stratigrahic horizons marked by powdered paint. The snow surface position (open circles) has been determined from data collected at the experimental site and augmented by data reported by the National Weather Service for the Fairbanks International Airport. Only significant snowfalls (s) are shown.

composed predominantly of depth hoar, sieving and stereological measurements made on photographs of disaggregated snow grains were acceptable methods of determining grain size and grain growth rates.

The results of the sieving are presented first in standard form to highlight the nature of the depth hoar as a geologic material. Figure 3-6 shows the cumulative curves of grain size distribution (in terms of ϕ-size, where $\phi = -\log_2$ [sieve mesh size (mm)]/$\log_2$ [1.0 mm] [Royse, 1970]) for three layers of snow on the ground, and two layers on the tables. At 0.04 m height above the ground, the snow was already depth hoar on November 28, when it was first sieved, so there was little further change in the grain size distribution during the winter. Higher (at 0.1 m and 0.2 m), large shifts in ϕ-size occurred as the snow coarsened, and the distribution changed from fine-grained and bimodal to coarse-grained and nearly unimodal. A distribution curve from January 15 at 0.3 m height for snow less than three days old represents the approximate starting configuration for the each layer and has been added to each set of curves for reference. Distribution curves for samples of snow on the tables, unlike snow on the ground, showed no systematic change, and maintained essentially the same distribution as new snow throughout the winter.

Following standard practice in sieve analysis, "grain size" was calculated from (Friedman and Sanders, [1983]):

$$\bar{d} = \frac{\sum_{j=1}^{L} M_j \bar{D}_{j,j+1}}{100} \qquad (3\text{-}1)$$

where $\bar{d}$ is the mean "grain size" of the sample, L is the number of

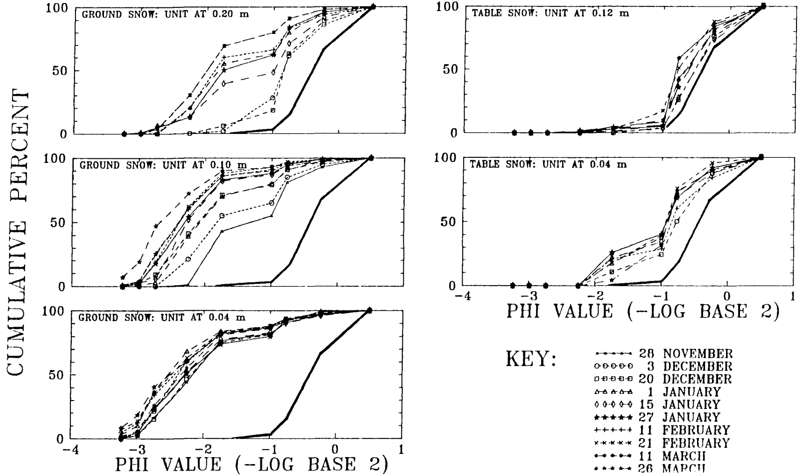

FIGURE 3-6: Cumulative size distribution curves for snow grains, 1987. See text for explanation. Phi Value = $-\log_2$(sieve mesh size [mm])/$\log_2$(1.0 mm). The cumulative curve for new snow is shown by a heavy line.

sieves, M_j is the weight fraction in the j^{th} sieve, and $\bar{D}_{j,j+1}$ is defined as the arithmetic mean of the diagonal of the sieve mesh through which the grain passed (sieve j+1) and the diagonal of the sieve mesh on which it was trapped (sieve j). The upper case D is used to clearly differentiate the size of the sieve mesh from the size of the grains measured by the sieves (lower case d). The results are plotted (Fig. 3-7).

"Grain size" was also measured from photographs using stereological techniques (see Section 2.2.3). In this case, the "grain size" plotted (Fig. 3-7) is the diameter of a circle with the same area as the mean cross-sectional area of the grain population (equivalent circle). The stereological and sieving values do not agree. The maximum difference occurs for snow which had metamorphosed into the largest, most ornate depth hoar grains, because these grains were least suited to characterization by an equivalent circle. For weakly metamorphosed snow (i.e., higher in the snow cover), the agreement is better. Both methods indicate that "grain size" increased as new snow metamorphosed into depth hoar.

Grain growth rates were easier to calculate and more accurately determined if grain mass rather than grain size was determined directly from the sieving. To do this, it was necessary to establish a relationship between sieve mesh size and the average mass of a grain trapped in that sieve. A sample of depth hoar was sieved and each grain contained in each sieve was weighed. For sieves with finer mesh, only a fraction of the total grains could be weighed and it was necessary to weigh 10 grains at a time, due to the resolution limits

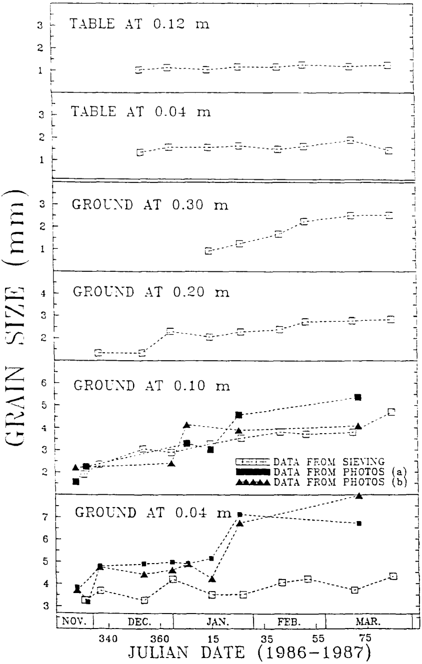

FIGURE 3-7: Snow "grain size" determined by sieving (open symbols) and photography (solid symbols). The two methods differed for snow grains which had metamorphosed the longest (0.04 m sample) and which had the most complex shapes. Two separate loose-grain samples ([a] and [b]) were photographed.

of the electronic balance. From the distribution of individual grain masses, the mean grain mass for a given sieve was determined. Within each sieve, the grain masses were generally normally distributed. These data and similar data reported by Bader et al. (1939) are plotted (Fig. 3-8). There is close agreement between the two data sets at the finer mesh sizes, but Bader's average grains weighed more for coarser meshes. This discrepancy probably results from differences in the way the sieves were shaken. Bader agitated his sieves for 40-60 minutes, vs. 30 seconds in this study. As a result, lighter grains were driven through the sieve mesh in Bader's case, producing a heavier mean grain in the residual.

The data were fit with a cubic polynomial (Fig. 3-8):

$$\text{(3-2)} \qquad \bar{m}_j = a_2(\bar{D})^2 + a_3(\bar{D})^3$$

where $\bar{m}_j$ is the average mass of an individual grain (kg) in the j^{th} sieve, $\bar{D}$ (equals $\bar{D}_{j,j+1}$ of Equation [3-1]) is the diagonal of the sieve mesh opening in mm, $a_2 = 1.149 \times 10^{-6}$ kg mm^{-2}, and $a_3 = 4.495 \times 10^{-9}$ kg mm^{-3} ($r^2 = 0.998$).

Using the polynomial fit, the average mass of a grain caught in a particular sieve could be determined from the sieve size. Since the total weight fraction of snow in the sieve was measured, the total number of grains in the sieve could be estimated. Summing for all sieves permitted the total number of grains in a sample to be estimated. The formulae and error estimates associated with these calculations are presented in Appendix IV.

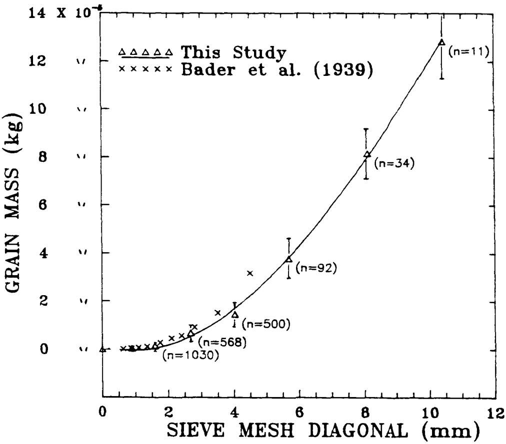

FIGURE 3-8: Snow grain mass as a function of sieve size. The number of grains sampled (n) at each sieve size is shown. A cubic polynomial was fit to the data (r^2= 0.998). Data from Bader et al. (1939) agree with the results of this study for the smaller sieve sizes, but diverge for the larger sizes due to differences in the length of time the sieves were agitated. Standard error of the mean is shown.

There was an order of magnitude decrease in the number of grains per unit volume as the snow on the ground metamorphosed into depth hoar (Fig. 3-9). Most of the reduction in number of grains occurred in the first few weeks after the snow was deposited. For example, samples from 0.04 m and 0.10 m contained more than 4 X 10^8 grains per cubic meter at the start of the winter; this was reduced to less than 5 X 10^7 grains per cubic meter in the first 40 days. During the next 110 days the number of grains hardly changed at all. Trajectories of number of grains per unit volume vs. time suggest that, regardless of when the snow started to metamorphose, the trajectories for all samples of snow on the ground were similar. In comparison, the number of grains per unit volume on the tables remained high throughout the winter and followed different trajectories. The reduction in the number of snow grains per unit volume during depth hoar metamorphism has been noted by many observers including Paulke (1934a), Seligman (1936), Bader et al. (1939), Benson (1962), and Akitaya (1974), but has not been quantified before.

The distribution of grain numbers as a function of mass (or sieve size) listed in Table 3-4 shows that the decrease in total grain number was accomplished by a decrease in the number of small grains and an increase in the number of large grains. This can be seen most clearly in samples taken from the snow on the ground at a height of 0.2 m. On December 3, 1986 (Julian Day 337) there were no grains of the four largest mass classes and there were over 10^8 grains per cubic meter in the smallest mass class. By the end of the winter, there were over 10^6 grains per cubic meter in the four largest mass classes

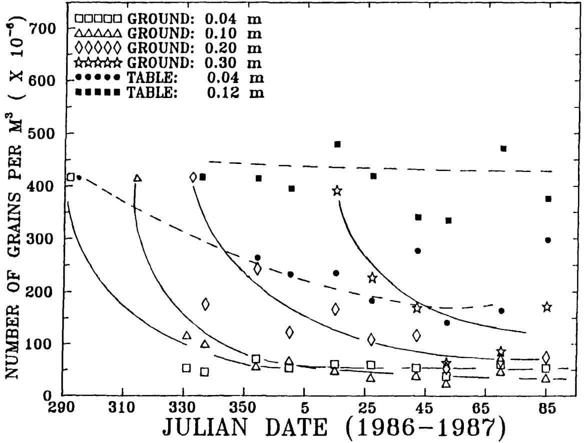

FIGURE 3-9: The number of snow grains per cubic meter, 1987. Trajectories of grain number vs. time have been drawn in by assuming that all samples had 400 X 10^6 grains per cubic meter when deposited; a value determined from snow samples at 0.2 and 0.3 m immediately after they were deposited. Trajectories for all snow layers on the ground were similar, but differed from the trajectories of snow on the tables.

TABLE 3-4: Number of grains (thousands) per cubic meter by sieve size, 1987.

DATE	8.7mm	7.3mm	5.7mm	4.2mm	2.8mm	1.9mm	1.4mm	0.9mm	0.3mm	TOTAL
SNOW ON THE TABLES AT 0.12 m										
354	0	0	0	0	0	1941	16581	90079	305421	414021
365	0	0	0	0	341	2153	31361	71937	289800	395592
15	0	0	0	0	249	1224	20242	90214	367779	479708
27	0	0	0	228	515	1162	28810	102718	285517	418950
42	0	0	0	0	1233	2102	26399	107694	203276	340704
52	0	0	0	0	822	8662	34781	88010	202759	335033
70	0	0	0	0	1136	3848	32877	101107	332828	471796
85	0	0	0	130	506	1322	53165	58460	262621	376204
SNOW ON THE TABLES AT 0.04 m										
354	0	0	0	0	2217	6469	20301	68446	166366	263798
365	0	0	0	0	4271	11492	32628	37934	146979	233305
15	0	0	0	0	4572	6270	28311	33008	163338	235499
27	0	0	0	0	5550	7084	22527	43992	104143	183296
42	0	0	0	0	4286	3852	24619	42979	201393	277128
52	0	0	0	0	3859	10889	29707	37678	59476	141608
70	0	0	0	0	10148	3818	13415	33029	104276	164686
85	0	0	0	0	1238	18886	47236	42466	188069	297896
SNOW ON THE GROUND AT 0.30 m										
15	0	0	0	0	45	409	6539	69466	315621	392080
27	0	0	0	0	0	5833	31691	41650	147931	227106
42	0	0	0	18	5953	4495	22445	37501	99434	169846
52	0	0	0	858	9664	7758	16033	16777	13241	64331
70	0	0	55	1480	13104	4841	10672	12408	44069	86629
85	0	13	459	2010	10444	6376	18339	24257	110641	172538
SNOW ON THE GROUND AT 0.20 m										
337	0	0	0	0	336	8081	16962	31189	119828	176396
354	0	0	0	0	1268	5727	35511	47525	152503	242536
365	0	0	0	2103	8426	4323	15547	31996	60510	122906
15	0	0	266	631	5246	3887	17489	35800	103697	167015
27	0	31	201	766	7988	5848	17729	22485	53241	108289
42	0	0	58	1754	8257	2896	12668	20835	70172	116640
52	0	0	206	2555	8638	5500	9094	13456	13655	53104
70	0	0	223	2443	11490	4323	8935	10388	29517	67320
85	18	35	361	1796	10028	4061	9735	12294	36041	74369
SNOW ON THE GROUND AT 0.10 m										
331	0	0	0	58	6828	4284	16225	15908	74000	117303
337	0	0	0	1712	6052	4255	13122	15885	59214	100241
354	0	59	268	2322	5033	2897	7403	11139	27745	56866
365	0	23	101	3306	6442	4321	8605	11612	34262	68671
15	0	79	122	3788	5604	2319	4532	6972	25338	48754
27	0	81	668	3055	5547	2159	4998	6641	11793	34942
42	14	92	998	2935	5187	1989	3507	5097	19310	39129
52	0	63	876	4225	5758	2500	3307	3845	4097	24671
70	22	92	1055	3448	5397	1949	3223	5621	26621	47429
85	141	363	1334	2207	3644	1479	2674	4086	18931	34859
SNOW ON THE GROUND AT 0.04 m										
331	0	62	481	1842	4893	1943	4735	7995	31655	53607
337	0	112	673	2969	3680	1970	3833	6387	26379	46003
354	0	65	526	2431	5090	2474	6135	9511	44676	70908
365	65	219	1329	2992	3350	1988	4570	7270	31779	53563
15	0	90	756	2022	4213	2028	4389	8363	39269	61131
27	18	100	736	2182	4039	2379	8522	9146	32483	59603
42	87	195	923	2042	4023	1882	3991	7078	33517	53738
52	116	164	847	1983	3311	1869	3708	5748	20414	38160
70	40	112	921	2547	5148	2235	4434	8592	36621	60650
85	159	277	979	1706	4151	1900	4355	7305	33021	53852

and the number in the smallest mass class had been reduced by a factor of 3. This contrasts with snow on the tables, for which there was no systematic change in the distribution of grains.

3.1.3 Depth Hoar Texture

The extreme temperature gradients found in the Fairbanks snow cover (Fig. 1-1) produce several metamorphic textures in the snow by late winter. These textures were observed from 1982 to 1989, and were photographed during the winters of 1986 and 1987 using the techniques explained in Section 2.2.2. C. S. Benson, who has studied snow in Fairbanks since 1962, confirmed that the snow was typical during the years the observations were made.

Five distinct textures, found as a sequence of metamorphic layers (Fig. 3-10), were identified based on the following criteria:

1) grain size,
2) crystal habit and aspect ratio of a:c axes,
3) c-axis orientation,
4) secondary crystallographic features such as striae, scrolling and eroded crystal edges,
5) qualitative "strength" of the snow.

Each texture was associated with a layer. These are described starting from the top of the snow cover. To facilitate discussing the textures, they have been called Metamorphic Layers 1 through 5. This was also done to emphasize that the textures developed as a metamorphic sequence which produced a layered structure in the snow.

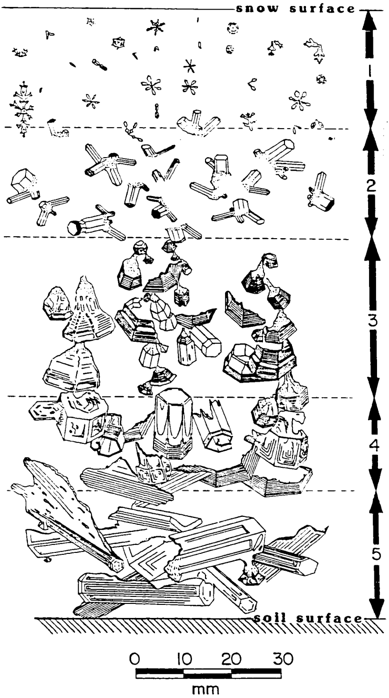

FIGURE 3-10: Five metamorphic layers found in the Fairbanks snow in late-winter. Prominent textural features which define each layer are shown and described in the text. Grain sizes are to scale, and layer thicknesses are shown in relative proportion.

The most intense metamorphism was found in the oldest layers at the base of the snow cover. Layer descriptions are cross-referenced to existing classification systems for snow flakes, and for snow which has been deposited on the ground. "ML" indicates the snow flake classification system of Magono and Lee (1966), which is similar to the classification system of UNESCO (1970), but uses different code letters. "SL" indicates the deposited snow classification system of Sommerfeld and LaChapelle (1970), and "C" indicates the deposited snow classification system devised by Colbeck (1986).

Metamorphic Layer 1 (Fig. 3-11) consists of new snow and snow showing slight cataclasis and some effects resulting from wind or above-freezing air temperatures (SL: III-A-1; C: II-C-1).

Metamorphic Layer 2 (Fig. 3-12) consists of small (≤ 1 mm) grains composed of squat, euhedral, hexagonal prisms with a:c axial ratios of about 1:1. C-axes orientations are random. Eugster (1950) called this type of snow crystals "prismatische Vollformen", and Akitaya (1974) called them "Ko-shimo-zapame-yuki" or solid-type depth hoar (SL: III-A-2; C: II-C-2; ML: Cle, Clg).

Metamorphic Layer 3 (Figs. 3-13a) consists of moderate sized grains (≤ 10 mm) composed of hexagonal, pyramidal cups which open downward and have a:c axial ratios of 2:1 or greater. The crystals are heavily striated, with sharp downward-facing edges. C-axes of the crystals generally fall within 15° of vertical, and the crystals are bonded into vertical ice columns with intervening, vertically elongated air spaces (Trabant and Benson, 1972), or "chains of grains" (Colbeck, 1986)(Fig. 3-13b). Bonds between crystals consist of

FIGURE 3-11: Snow grains from Metamorphic Layer 1. Partial and nearly complete dendrites and spatial dendrites are present. Crystal edges are smooth and rounded suggesting that vapor transfer is taking place.

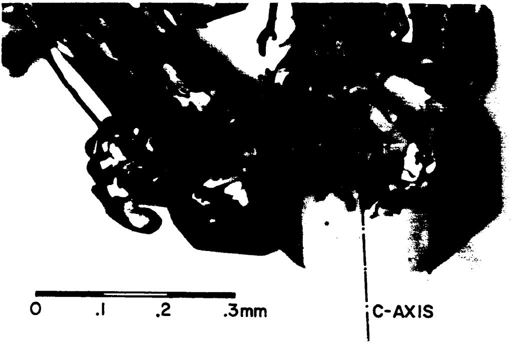

FIGURE 3-12: Snow grains from Metamorphic Layer 2. Crystals are solid, euhedral, hexagonal prisms known as solid-type depth hoar. Crystal edges are sharp and striae are absent. C-axes orientation are random.

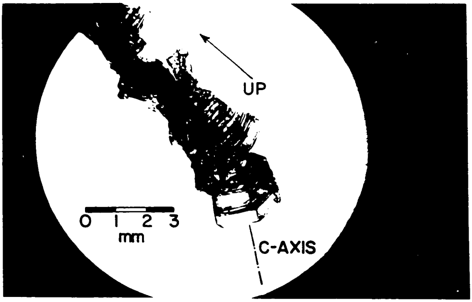

FIGURE 3-13a: A chain of cup crystals from Metamorphic Layer 3.

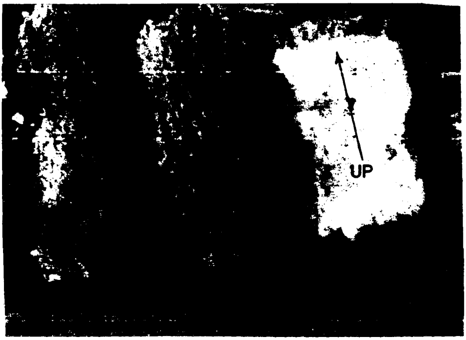

FIGURE 3-13b: Snow samples from the Brooks Range, Alaska, showing the vertical structure of Metamorphic Layer 3. Wind incorporated dirt and dust in the snow during its deposition, highlighting the texture. (photo by C. S. Benson).

unfaceted ice attached to the crystals like a calyx attaches to a flower. The bonding is weak, making the layer fragile. Eugster (1950) called the type of crystals found in this layer "Hohlformen Basale Teilbecher", and Akitaya (1974) called them "Shimo-zarame-yuki" or skeleton-type depth hoar, but neither author reported the vertical chains of grains (SL: II-B-2; C: II-C-3; ML: clh).

Metamorphic Layer 4 (Fig. 3-14) consists of large (≤ 20 mm) grains which show evidence of erosion by sublimation. The grains are aggregates of several crystals which are either hollow, hexagonal columns with axial ratios (a:c) of about 1:2, or partial crystals whose interfacial angles suggest that they are the remnants of cup-like crystals which were greater than 20 mm across. C-axes of both types of crystals are approximately vertical. Downward-facing crystal edges are sharp, and have pronounced striae. Several features found on the upper parts of crystals (rounding, ragged edges, loss of striae, and smooth crystal faces, sometimes with irregular holes) indicate erosion by sublimation (Colbeck, 1986). Bonding between grains is poor. In fact, Metamorphic Layer 4 can be identified when sampling for density, because it is the least cohesive layer, and it is difficult to sample. The hollow columns found in the layer are similar to ML type Clf, and Eugster (1950) called similar crystals "Hohlformen prismatische becher (Not included in the SL or C classification system).

Metamorphic Layer 5 (Fig. 3-15) consists of grains which average 10 to 20 mm in length, though grains longer than 30 mm can be found. Most grains consist of several elongated, hexagonal column crystals

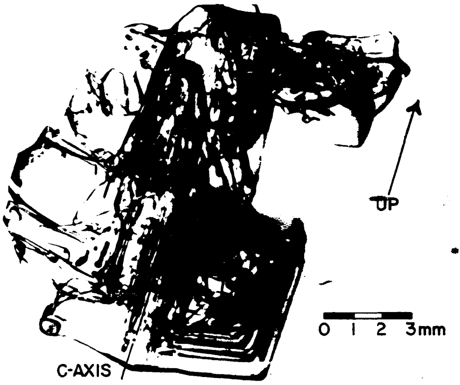

FIGURE 3-14: A column crystal from Metamorphic Layer 4. The crystal exhibits the distinctive erosional features such as rounded upper edges, glassy crystal surfaces, and truncated crystal facets which are typical of this layer. C-axes are usually vertical.

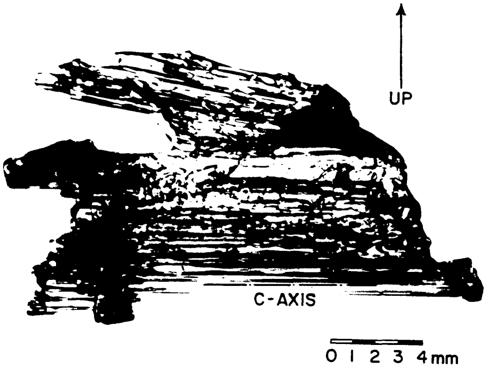

FIGURE 3-15: A grain from Metamorphic Layer 5. It consists of several elongated prismatic columns with keel-like growths on their upper surfaces. Lower edges are sharp, indicating active crystal growth. A small "secondary" cup crystal is growing downward from the bottom edge of the largest column. C-axes orientation of most large columnar crystals are horizontal to sub-horizontal.

with axial ratios (a:c) of approximately 1:5. C-axes of most crystals are close to horizontal. Crystals are euhedral, and are generally more complete than those of Metamorphic Layer 4 and erosion features are not prominent. The majority of crystal faces show skeletal growth (i.e., they step inward towards the c-axes [Knight and DeVries, 1985]). Occasionally "secondary" growth features, such as downward-facing cup-crystals, with vertical c-axes, grow like small buds along the bottom edges of large columns. The bonding of the layer is surprisingly strong, particularly in contrast to the overlying layer, producing a felt-like texture of interlocking horizontal crystals. The column crystals are similar to the "needle" and "sheath" crystals photographed by Akitaya (1974) (Not included in the SL or C classification systems).

The five metamorphic layers developed every winter during which observations were made, though the relative thicknesses of each layer varied from winter to winter. The thicknesses of Metamorphic Layers 1, 2, and 3 were difficult to determine because the boundaries between these layers were gradational. The lower boundary of Metamorphic Layer 3, however, could be located by gently scraping a snow pit wall, because the cohesionless crystals of Metamorphic Layer 4 were removed, leaving the "chains of grains" in Metamorphic Layer 3 intact. The boundary between Metamorphic Layers 4 and 5 was usually distinct because of the change in c-axes orientation. This boundary swept upward through the winter. In December 1987, Metamorphic Layer 5 was 20 mm thick, by March it was 80 mm thick.

A reconnaissance survey conducted during this study suggested that Metamorphic Layer 5 was better developed (i.e., larger column crystals) in snow deposited on lake or river ice than over bare soil, and it appeared to be better developed over moss tussocks than other vegetation. A qualitative comparison of two snow pits, one over bare soil and the other over an impermeable tarp, indicated that Metamorphic Layer 5 was better developed over the bare soil, but was not absent over the tarp.

Metamorphic Layers 1 and 2 are common in most snow covers. The cup crystals of Metamorphic Layer 3 have been described by Paulke (1934a,b), Seligman (1936), Eugster (1950), Sommerfeld and LaChapelle (1970), Trabant and Benson (1972), Akitaya (1974), Pahaut and Marbouty (1981), Kolomyts (1984), and Colbeck (1986), but the vertical structure (Fig. 3-13b) has received less attention. Kojima (1956), working on snow with a texture similar to Metamorphic Layer 3, found that depth hoar was "stiff" in vertical compression, but subject to brittle failure. Akitaya (1967, 1974) suggested that a "vertical columnar structure" was the cause of this property. Benson and Trabant (1972) and Trabant and Benson (1972) noted "vertically oriented channels" in the Fairbanks snow. Sommerfeld and LaChapelle (1970) observed a texture they called "late-stage advanced temperature gradient metamorphism" or "lattice grains" which is probably equivalent to Metamorphic Layer 3. Most recently, Colbeck (1986) described one aspect of the texture as "chains of grains".

Metamorphic Layers 4 and 5 have not been described before, perhaps because most snow covers do not metamorphose long enough or under strong enough temperature gradients to develop the five-layer sequence. Eugster (1950) identified a metamorphic sequence in which solid depth hoar crystals developed into skeletal crystals or the reverse. Sommerfeld and LaChapelle (1970) identified a similar sequence of early- and late-stage depth hoar metamorphism, with their late-stage corresponding with Metamorphic Layer 3. Akitaya (1974) also described a sequence which includes Metamorphic Layers 1 through 3. Bradley et al. (1977a,b) and Adams and Brown (1982a) associated changes in the strength of snow layers with the development of depth hoar. They found minimum strength in a snow layer consisting of small, anhedral grains, probably equivalent to the transition between Metamorphic Layer 1 and 2, while in this study, the minimum strength was observed in Metamorphic Layer 4.

The transitions in crystal habit or form observed in the five metamorphic layers can be explained, in part, by differences in the local temperature and the temperature gradient with height in the snow, since the base of the Fairbanks snow cover is often -2°C while the top is -40°C. In clouds, snow crystal habit is determined by temperature, with secondary features determined by growth rate or the supersaturation at the crystal surface (Nakaya, 1954; Kobayashi, 1961; Mason et al., 1963; Magono and Lee, 1966; Lamb and Hobbs, 1971; Lamb and Scott, 1972; Frank, 1982; Keller and Hallett, 1982). Distinct and abrupt transitions between hexagonal plate crystals and hexagonal columns occur at -4°, -9°, and -22°C. Akitaya (1974; p. 28) found that similar transitions occurred for snow deposited on the ground, though he related the transitions to temperature and temperature gradient. Marbouty (1980) and Adams and Brown (1982b) have reported similar results. However, the c-axes orientations observed in three of the five metamorphic layers cannot be explained by changes in temperature or temperature gradient. A speculative explanation for the c-axes orientation, based on the convective circulation pattern, is discussed in Section 5.4.

3.2 Measurements Related to Heat Transport in the Snow

3.2.1 Heat Flow

Heat flow at the snow-soil interface, measured using thermopile heat flow meters (HFM's)(see Section 2.3.1), varied between 5 and 30 W m^{-2} during the winters of 1985 to 1987 (Fig. 3-16). In comparison, the geothermal heat flow is only about 0.05 W m^{-2}. The major source of the heat came from latent heat liberated as the soil moisture froze. This can be verified by comparing the latent heat flow released in 1987 to the heat flow measured by the heat flow meters. The latent heat flow, calculated from the soil moisture profiles (see Section 2.4.1) and the rate at which the the freezing front moved downward in the soil (see Section 2.3.3), varied between 6 and 24 W m^{-2}, which is in broad agreement with the heat flow measured by the HFM's.

There was a marked difference in the heat flow signal between the winter of 1985 and the next two winters. The 1985 signal varied smoothly and regularly. The mean heat flow was approximately 6 W m^{-2}, except during two discrete periods when air convected in the snow (see

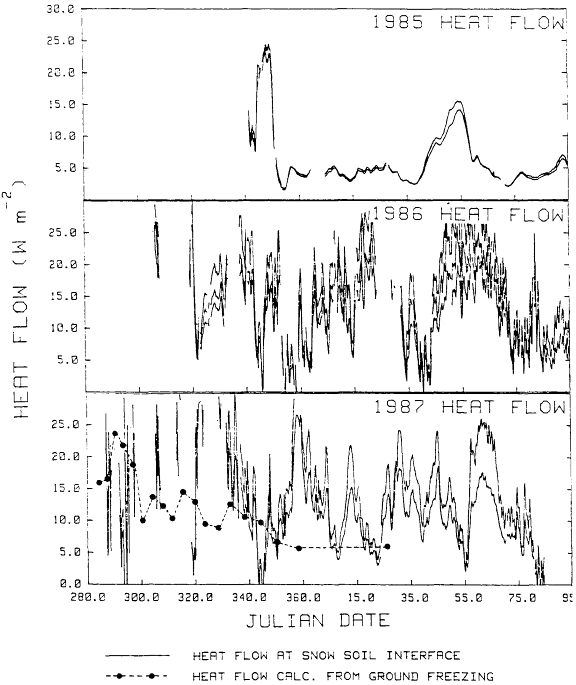

FIGURE 3-16: Heat flow at the snow-soil interface, 1985-1987. Heat flow was measured by heat flow meters (see Section 2.3.1) and from soil temperature and moisture data.

Section 5.1). During the next two winters the heat flow showed greater and more rapid fluctuations, and its mean value was about 10 W m^{-2} (Fig. 3-16).

The interannual differences in heat flow were the result of differences in the snow depth during 1985 and in the next two winters (Fig. 3-17). The thinner snow of 1986 and 1987 gave rise to larger and more rapidly varying temperature gradients than those which were present in 1985. The larger temperature gradients and fluctuations of the gradients resulted in greater and more rapidly varying heat flow.

3.2.2 Thermal Conductivity

The bulk effective conductivity of the snow can be written as:

$$k_{eff} = Q_s / \overline{[\partial T/\partial z]} \tag{3-3}$$

where Q_s is the heat flow at the snow-soil interface, and $\overline{\partial T/\partial z}$ is the average vertical temperature gradient across the snow cover. It showed little difference from one winter to the next, and ranged from 0.1 to 0.3 W $m^{-1}K^{-1}$ (Fig. 3-18).

Sporadic needle probe measurements of the thermal conductivity of the snow in 1985 became routine in 1986 and 1987 (see Section 2.3.4). The measurements showed that the thermal conductivity of depth hoar was lower than would be predicted from its density. Most authors (Hjelström, 1890; Jansson, 1901; Okada, 1905; Devaux, 1933; Kondrat'eva, 1954; Yosida and Iwai, 1954; Mellor, 1977) suggest that

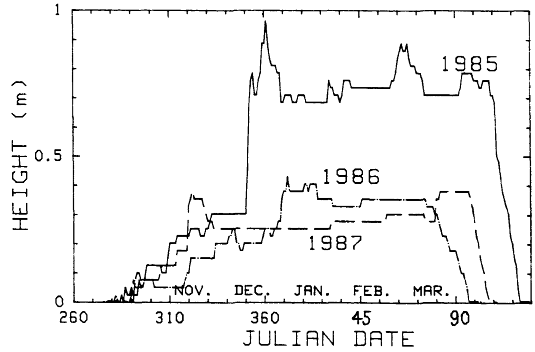

FIGURE 3-17: Fairbanks snow depth, 1985-1987. The large snowfall in December, 1984 was the single largest snowfall for Fairbanks in 10 years (see also Fig. 1-3).

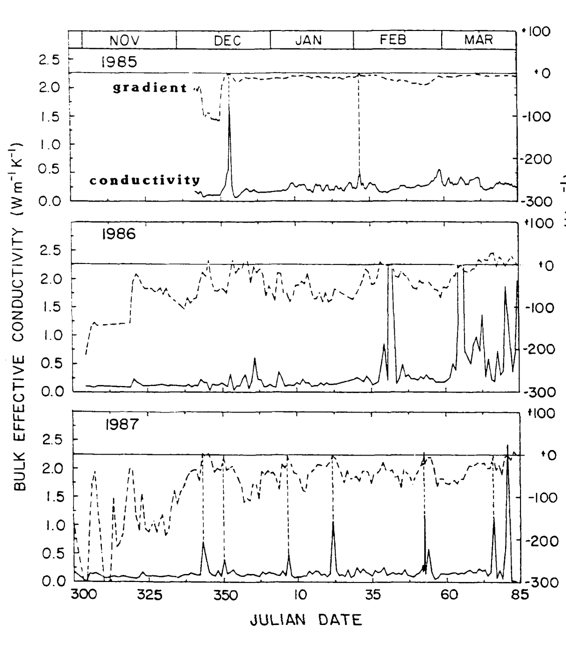

FIGURE 3-18: The bulk effective thermal conductivity of the Fairbanks snow cover, 1985-1987. The conductivity (solid line) was calculated by dividing the heat flow at the snow-soil interface (Fig. 3-16) by the average vertical temperature gradient (dashed line). When the temperature gradient approached zero, the conductivity was undefined, producing large spikes. The mean value of the conductivity was 0.1 to 0.2 W $m^{-1}K^{-1}$ for all three winters.

for snow with a density of 200 kg m^{-3}, typical of depth hoar, the thermal conductivity (in the absence of convection) would range from 0.10 and 0.20 W $m^{-1}K^{-1}$, whereas the measured values ranged from 0.04 to 0.10 W $m^{-1}K^{-1}$ (Appendix III). It is likely that the thermal conductivity values determined in this study were lower than predicted because the predictive equations were based on studies which did not include measurements of depth hoar. However, few of the studies reported the type of snow or the temperature at which measurements were made, and because thermal conductivity is a function of temperature and snow texture as well as density (de Quervain, 1972), the actual cause of the discrepancy cannot be determined. Lange (1985) measured thermal conductivity of depth hoar and found values of 0.04 to 0.05 W $m^{-1}K^{-1}$, which is within the range determined in the present study.

The bulk effective thermal conductivity of the snow (Fig. 3-18) was higher than the average values measured by needle probe. The disagreement between these two values is discussed in Section 5.1.2 as evidence for convection in the snow.

The effective thermal conductivity (k_{eff}) was measured over a range of temperatures to determine its temperature dependence. Unfortunately, the range of the circulating bath (see Section 2.3.4) limited normal measurements to -20° to 0°C, and measurements near 0°C were difficult to make without melting the sample. To separate the temperature effects from the effects of snow density and texture, the results for each snow sample were plotted separately (Fig. 3-19a). Over the range of temperature for which measurements were made, the

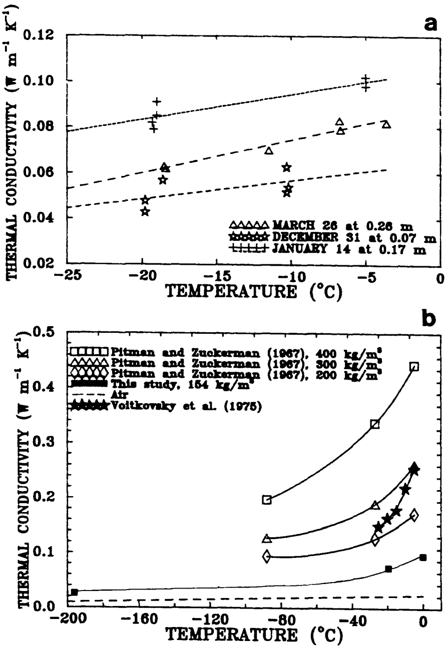

FIGURE 3-19a: The temperature dependence of the thermal conductivity of depth hoar, 0° to -20°C. Different textures are plotted separately.

FIGURE 3-19b: The temperature dependence of the thermal conductivity of depth hoar, 0° to -196°C. Data for dry air (dashed line), from Pitman and Zuckerman (1967), and Voitkovsky et al. (1975) are included for comparison.

thermal conductivity was linear and the slopes of the lines were similar, suggesting that $\partial k/\partial T$ may be independent of snow texture.

In order to get a datum at a temperature for which there would be no latent heat transfer, one measurement was made with the sample jacketed and immersed in liquid nitrogen (-196°C). Using measurements made at -5°, -20°, and -196°, and with the thermal conductivity of air as a limiting value, a composite curve for the temperature dependence of k was plotted (Fig. 3-19b). Data for other snow, measured by Pitman and Zuckerman (1967) and Voitkovsky et al. (1975), are plotted for comparison. The methods used in the latter study were not described.

During 1987, the thermal conductivity of three layers was measured throughout the season (Fig. 3-20). The layer at 0.07 m height in the snow was already completely metamorphosed into the texture described as Metamorphic Layer 3 when the initial measurement was made (see Section 3.1.3). As the layer metamorphosed into Metamorphic Layer 4 and then Metamorphic Layer 5, its thermal conductivity first decreased then increased. This is consistent with the observations of the relative strength and cohesiveness of the metamorphic layers. Metamorphic Layer 4 is the least cohesive, most poorly bonded layer, which would contribute to the low thermal conductivity. Metamorphic Layer 5 is stronger, more cohesive, and has a higher thermal conductivity than Metamorphic Layer 4.

At 0.16 m height, the initial measurement was in new snow. Compaction of this very light snow caused an increase in thermal conductivity, but as Metamorphic Layer 3 and Layer 4 textures

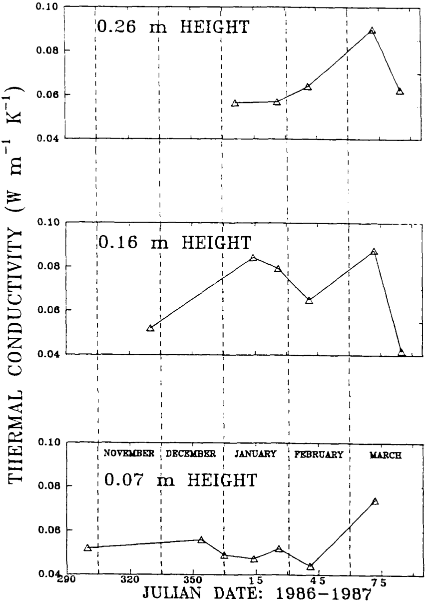

FIGURE 3-20: Changes in the thermal conductivity of three snow layers, 1987. The layer at 0.07 m was already depth hoar when first measured.

developed, the thermal conductivity decreased. Results for the stratum at 0.26 m were similar, but metamorphism did not progress as completely as in the underlying strata.

3.2.3 Temperature

The fluid in a porous medium which is heated from below and cooled from above will convect when buoyancy forces overcome the viscous resistance of the fluid. When this occurs, regions of upwelling warmer fluid and downwelling cooler fluid will develop. The thermistor arrays (see Section 2.3.2) used in this study to measure in-situ snow temperatures were designed to detect the horizontal temperature gradients resulting from fluid (air) flow in the snow. In the absence of convection, the horizontal plane defined by the thermistor strings was expected to be isothermal because the snow layers were horizontal and nearly homogeneous.

The horizontal plane was rarely isothermal during all three winters. At times the temperature along horizontal thermistor strings varied as much as 16°C over distances of about a meter. To illustrate these large temperature variations, the temperature deviation, defined as the difference between the observed thermistor temperature and the mean temperature of the entire string, is plotted in Figure 3-21 for a selected horizontal thermistor string from each winter.

The large temperature deviations shown in Figure 3-21a were first reported for the winter of 1985 by Johnson et al. (1987), who were able to show that the deviations could only be accounted for by

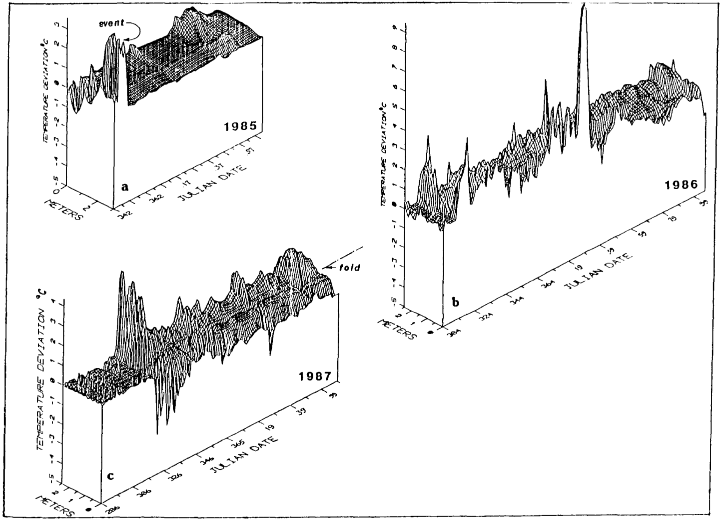

FIGURE 3-21: Temperature deviations for the horizontal thermistor strings, 1985-1987. The flat part of the surface from early-winter, 1986 and 1987, results from when the thermistors were still in the air. The fold-like feature in 1987 is the result of one displaced thermistor. The "event" of 1985, discussed in the text, is marked.

convection in the snow. Temperature deviations were present more frequently during the next two winters, indicating that convection was present in the snow cover during all three winters that the study was in progress. The two horizontal strings installed in 1985 (see Section 2.3.2 and Fig. 2-2) were insufficient to allow the geometry of the convective flow to be deduced for that winter. The increased number of thermistor strings used in 1986 and 1987 (Fig. 2-3) made it possible to infer some aspects of the convective circulation which are discussed in Section 5.2.1.

The magnitude and prevalence of temperature deviations differed between 1985 and the next two winters (Fig. 3-21), due largely to differences in the snowfall history (Fig. 3-17). The temperature deviations of the large event in the winter of 1984-85 virtually vanished on December 17, 1984 (Julian Day 351) when 0.45 m of snow fell in an unusual snow storm, as discussed by Johnson et al. (1987). New snow and high air temperatures reduced the vertical temperature gradient by a factor of 20 and compressed the initial snow layer, which was already depth hoar, to slightly more than half its original thickness, greatly reducing its permeability. During the following two winters there was only half as much snow and no large snowfalls.

The temperature records from the horizontal thermistor strings could be classified into one of three types of events: 1) isothermal, 2) coherent, or 3) incoherent. Isothermal events were periods when all thermistors on a string registered temperatures within ± 0.2°C of the mean temperature (Fig. 3-22a). Coherent events were periods when temperature records were synchronous, and the range of temperatures

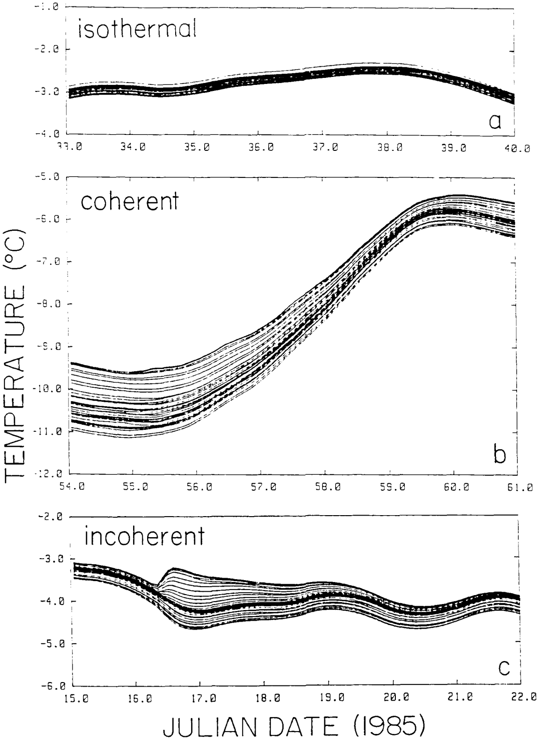

FIGURE 3-22: Three types of temperature events.

about the mean was greater than $\pm$ 0.2°C. The temperature of all thermistors rose and fell simultaneously in these events so that their traces were parallel and did not cross (Fig. 3-22b). As mean temperature rose and fell, the range of temperatures decreased and increased. Incoherent events were periods when temperature records were asynchronous and the temperature of some thermistors decreased while others increased, resulting in the crossing of thermistor traces (Fig. 3-22c). The transition between isothermal and coherent events was gradational, based solely on the range of temperature, but the transition from isothermal or coherent events to incoherent events was abrupt (Fig. 3-22c). In most cases, incoherent events ended when they gradually changed into coherent events. Incoherent events were not due to the spurious behavior of one or two thermistors; they reflect deviations in temperature throughout the entire horizontal thermistor plane. For example, during one incoherent event in 1987, 45 thermistors warmed at rates up to 0.05° hr^{-1}, while 111 thermistors cooled at rates up to 0.10° hr^{-1} (Fig. 3-23).

Coherent events were present about 20% of the winter of 1985. In 1986 and 1987, with the snow cover half as thick, they were present virtually throughout the winter except when incoherent events occurred or during the few times the temperature gradient approached zero. Incoherent events were present 15% of the winter of 1987, 16% of the winter of 1986, and 8% of the winter of 1985. During coherent events, the range of temperature on the horizontal strings often exceeded 5°C and occasionally reached 10°C. During incoherent events, the range was greater, exceeding 16°C at one time.

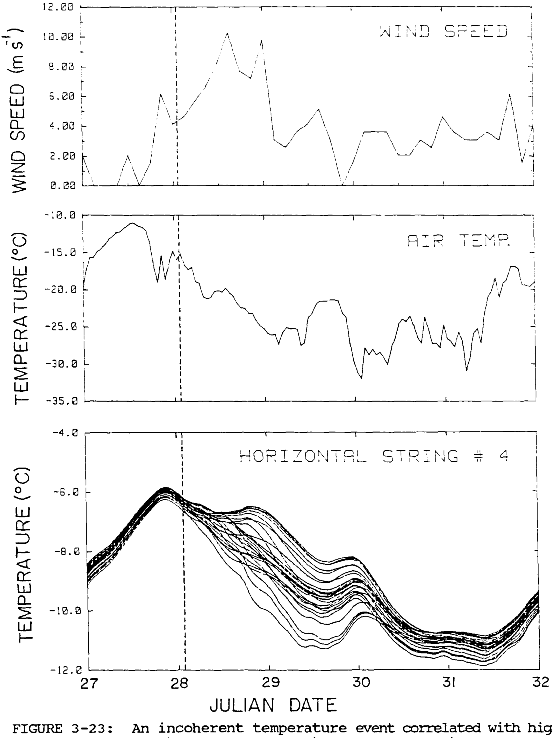

FIGURE 3-23: An incoherent temperature event correlated with high wind speed. In this event 45 thermistors were warming, 111 were cooling (for all 6 horizontal strings).

Eleven of the 25 incoherent events identified between 1984 and 1987 correlated with the few periods when the wind exceeded 5 m s^{-1} (Table 3-5). Theoretical studies (Clarke et al., 1987; Colbeck, in prep.) have suggested that the temperature in the snow could be affected by wind, but this is the first demonstration of a close correlation. One example of the correlation between an incoherent event and high wind is shown in Figure 3-23. This event, the most conspicuous incoherent event of 1987, began when the wind speed increased rapidly to the highest value recorded during the winter.

Four of the 14 incoherent events not correlated with wind occurred 10 to 20 hours after the air temperature dropped sharply. If it is assumed that the events were triggered by the air temperature, then the onset of the events occurred 24 to 48 hours before the cold wave could have penetrated to the depth of the thermistors by diffusion. For example, if we define the start of an incoherent event as the onset of rapid divergence in the temperature of the thermistors, event B of Figure 3-24 started 17 hours after the air temperature dropped sharply. At that time, the thermistors were covered by 0.13 m of depth hoar with a thermal diffusivity estimated to be $10^{-7}m^2$ s^{-1}. If the drop in temperature propagated into the snow in a purely diffusive manner, the time necessary for it to reach the thermistors would have been 40 to 50 hours (Turcotte and Schubert, 1982), rather than 17 hours which was observed. This point is discussed further in Section 5.1. The other incoherent event labeled A in Figure 3-24 is typical of the events which correlated with neither strong wind nor rapid changes in air temperature.

TABLE 3-5: Correlation of wind with incoherent events.

EVENT	JULIAN DATE	AVE. WIND SPEED ($m\ s^{-1}$)	PEAK GUST ($m\ s^{-1}$)	CORRELATED
		1984 - 1985		
1	345	0.0	0.0	
2	16	4.0	7.0	*
3	45	6.9	10.1	*
4	107	---	9.0	*
NONE	351	---	8.0	
NONE	62	---	3.0	
		1985 -1986		
5	320	0.6	3.0	
6	331	1.3	3.5	
7	---	---	---	
8	351	5.0		*
9	357	7.0	11.5	*
10	---	---	---	
11	8	1.6	3.2	
12	11	4.1	6.0	*
13	18	3.5	5.5	
14	21	1.9	2.8	
15	28	5.2	6.9	*
16	34	---	13.4	*
NONE	40	---	5.6	
		1986 - 1987		
17	323	5.5	---	
18	326	---	---	
19	335	4.5	---	
20	351	5.0	---	*
21	363	5.0	---	*
22	28	10.0	---	*
23	55	6.1	---	*
24	58	2.8	---	
25	63	1.0	---	

--- indicates no reliable data
* indicates event correlates with wind

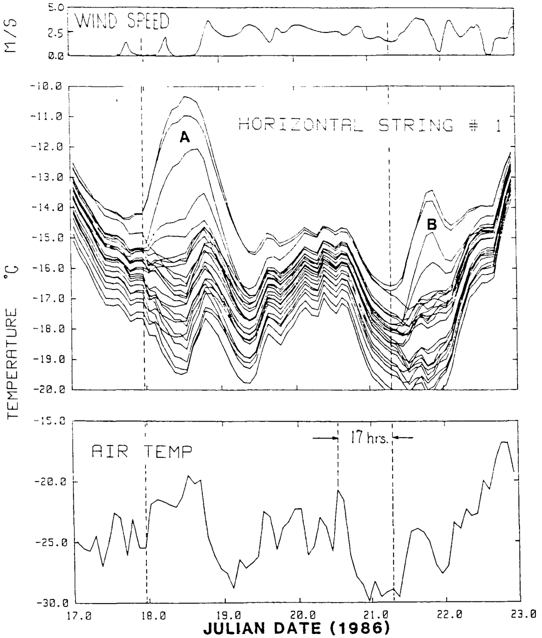

FIGURE 3-24: Two incoherent temperature events, one correlated with a sharp drop in air temperature. Event A does not correlate with wind speed or changes in air temperature. Event B began 17 hours after a distinct drop in air temperature. Snow thickness above thermistors was 0.13 m.

During coherent and incoherent events, relatively warm and cold zones developed in the snow and persisted throughout most of the winter, evolving slowly, but never changing location abruptly (Fig. 3-25). The two horizontal strings used in 1985 were insufficient to determine the presence of these zones, but in both 1986 and 1987 their presence was definite. For example, in 1987 a relatively warm zone developed on the north edge of the thermistor plane and persisted for more than a month (Fig. 3-25b to 3-25d) before it evolved into two warm zones (Fig. 3-25e). The zones diminished or disappeared only during periods when the air temperature increased and the vertical temperature gradient in the snow approached zero, as can be seen in Figure 3-25i. The warm and cold zones reappeared in approximately the same locations following the periods of higher air temperatures.

A correlation analysis was used to verify that the relative warm and cold zones stayed in the same location and maintained the same spatial relationship to each other. The analysis was done by selecting an arbitrary reference date for each winter and comparing the locations of the temperature maxima and minima for all other days of the winter to their locations on the reference date. An auto-correlation function (Davis, 1973) was used; correlation coefficients (r-values) near 1 implied that temperature maxima and minima remained in nearly the same location as on the reference date.

Johnson et al. (1987) performed separate correlation analyses for each horizontal string used in 1985, and found that the warm and cold spots remained in fixed locations during the 10-day event which ended on December 18, 1984 (Julian Day 353), but after that, the r-values

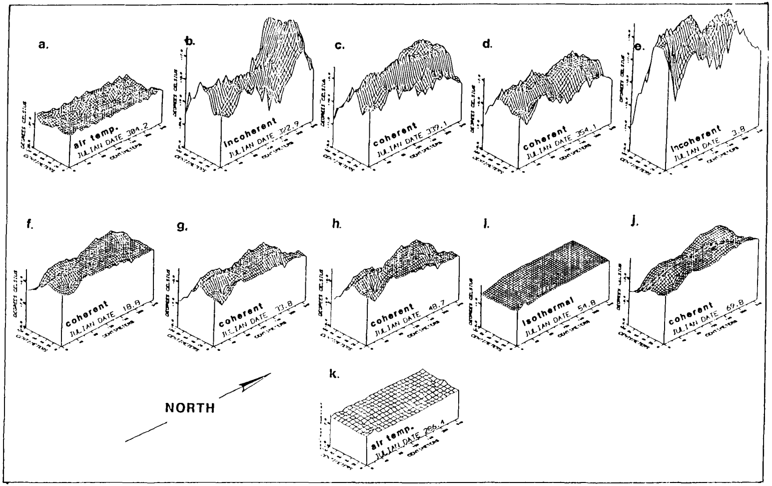

FIGURE 3-25a-j: The evolution of relative warm and cold zones in the horizontal temperature field, 1987. Temperature is the vertical coordinate; increment is 1°C. Higher temperatures appear as peaks, lower temperatures as valleys. Surfaces of little relief correspond with when the thermistors were not yet buried in snow or when the vertical temperature gradient approached zero.

fell from 0.9 to near 0, implying that through the rest of the winter the location of warm and cold spots was random (Fig. 3-26a). For 1986 and 1987, all six horizontal thermistor strings were treated as a single composite string consisting of 154 thermistors. The analysis showed that during each winter the warm and cold spots persisted in the same location throughout the winter (Figs. 3-26b, 3-26c). For example, the temperature field from December 21, 1986 (Julian Day 355) was chosen as the reference date for 1987. At the beginning of the winter, when the thermistors became buried in snow, there was a rapid rise in the r-values from approximately 0 to greater than 0.7. During the entire period that the strings were covered in snow the r-values were high. The three brief exceptions, when r approached 0, coincided with the three thaws during which the vertical temperature gradient changed sign (Fig 1-1).

A second correlation analysis was used to check the possibility that the spatial persistence of warm and cold zones during 1986 and 1987 might have been the result of vertically displaced thermistors or variations in the snow cover thickness. Before being covered by snow, the thermistors were within ± 2 mm of their specified vertical positions. Excavation of the thermistors after they had settled through the winters indicated that the maximum vertical displacements from the mean height were less than ± 10 mm for all but one thermistor in 1987 (Fig.3-27c), which, combined with spatial variations in the snow depth (Fig. 3-27a), produced variations in the snow thickness above each thermistor of ± 10 mm (Fig. 3-27b). Correlation of the temperature deviations at each thermistor with a) thermistor height,

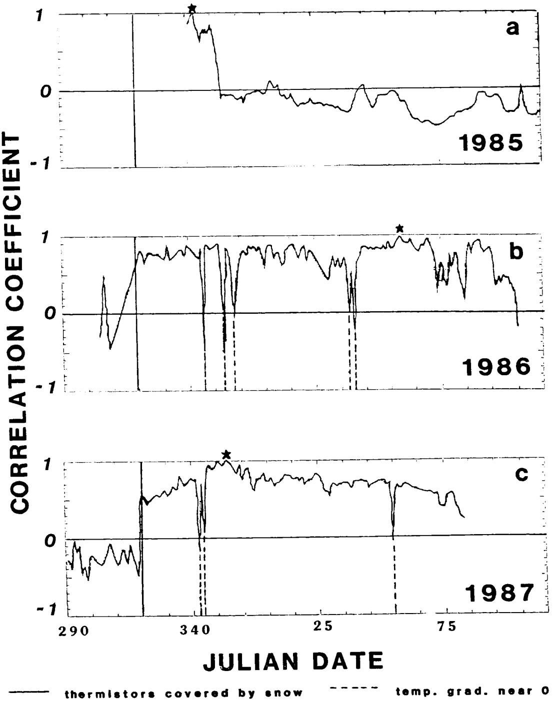

FIGURE 3-26a-c: Correlation between the horizontal temperature field on a reference day and the rest of the winter, 1985-1987. High correlation (≈1) indicates that warm and cold zones stayed in the same location. * indicates the correlation base date.

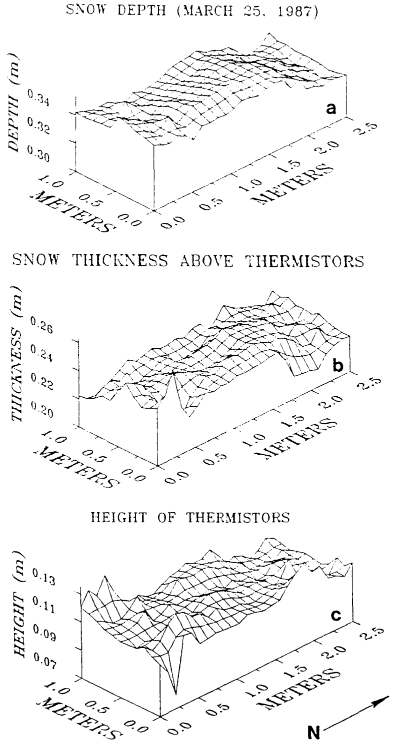

FIGURE 3-27a-c: Thermistor heights and snow depth, 1987. Thermistors were excavated on March 25, 1987; one thermistor located in the SE corner was found displaced downward 20 mm from the mean height.

b) the snow layer thickness above each thermistor, and c) the total snow depth at each thermistor indicated that the best correlations were achieved for b), but even in this case r-values averaged less than 0.40. Thus, vertical displacement of the thermistors or variations in the snow thickness could not explain the persistence of the warm and cold zones.

The persistence of warm and cold zones in the snow suggested that there were spatial variations in the temperature at the base of the snow, so in 1987 a string of five thermistors was placed at the snow-soil interface (see Section **2.3.2** and Fig. 2-3). Interface temperatures were never isothermal, with horizontal temperature differences exceeding 4°C early in the winter (Fig. 3-28). Measurements made by Gosink et al. (1988) in the same field confirm that differences of several degrees exist at the snow-soil interface every winter.

Regions of higher temperature at the snow-soil interface generally coincided with relative warm zones in the snow. This correlation could only be made along the southern edge of the thermistor array where the snow-soil interface thermistors had been placed (Fig. 2-3), and unfortunately, the maximum temperature warm zone in the snow was along the north edge of the array (Fig. 3-25). However, along the southern edge, the higher temperatures in the snow coincided with the higher temperatures at the interface throughout the winter.

One possible reason for the variation in temperature at the snow-soil interface was that the soil was not uniformly saturated prior to

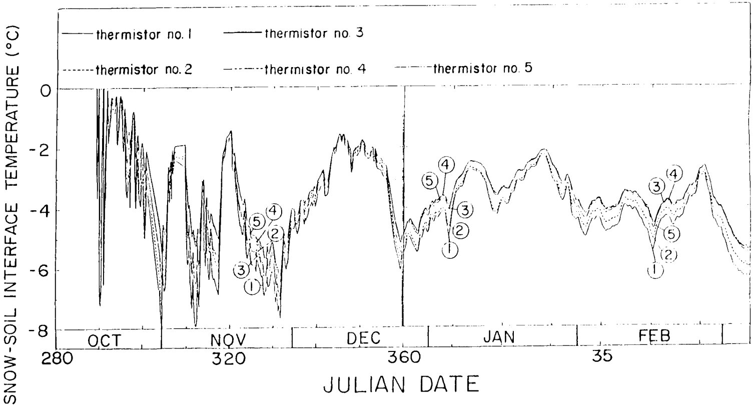

FIGURE 3-28: Temperature at the snow-soil interface, 1987.

the onset of freezing, resulting in an unequal release of latent heat. In order to verify that the soil surface temperature would differ between adjacent wet and dry spots, the Stefan Solution (Turcotte and Schubert, 1982) was used to calculate the snow-soil interface temperature as a function of time. It was assumed that both wet (50% moisture by dry weight) and dry (10% moisture) soils were silt covered by 0.5 m of snow with its upper surface maintained at a constant temperature of -20°C. The two soils were initially at 0°C throughout. Results indicate that after 30 days, the wet soil would be 2°C warmer at the surface than the dry soil. For an inexhaustible moisture supply, the temperature of the two adjacent spots would continue to diverge with time, but if all the available moisture froze, the two temperatures would begin to converge.

Temperature differences of more than 1° persisted at the snow-soil interface through March 6, 1987 (Julian Day 65) (Fig. 3-28), 110 days after latent heat effects should no longer have contributed to maintaining the temperature contrast because all the available moisture would have been frozen. Soil moisture measurements made at the experimental site indicate that below 0.4 m depth, the soil moisture content is negligible (≤ 16%). The freezing front reached this depth by November 16, 1986 (Julian Day 320). However, as the ground freezes, wet spots become ice-rich and dry spots become ice-poor. Ice-rich soil has a thermal conductivity four times higher than dry soil (Johnston, 1981), which is adequate to maintain the differences in snow-soil interface temperatures.

The vertical thermistor strings yielded data from which the temperature gradient, $\frac{\partial T}{\partial z}$, and curvature, $\frac{\partial^2 T}{\partial z^2}$, could be calculated. Negative curvature (concave-downward) persisted throughout the study, with only a few exceptions. A sequence of typical vertical temperature profiles from 1985 are shown in Figure 3-29. They represent most of the profiles measured during the study; few profiles were linear or concave upward.

The negative curvature of the profiles could have been caused by 1) the response of the snow to rapidly falling air temperatures, 2) vertical variation of the thermal conductivity, 3) latent heat liberated as vapor diffuses upward, and 4) vertical advection of warm air. By examining only periods of stable air temperature or periods when the air temperature fluctuated closely about a mean value, curvature due to rapidly falling air temperature (1) can be avoided. For example, the curves in Figure 3-29 were measured at a time when the air temperature fluctuated less than ±8° about a mean of -30°C, yet they are still strongly curved. The vertical variation of thermal conductivity (2) was not the cause of the vertical curvature because there was little variation in conductivity, and it would have given rise to positive rather than negative curvature (because the snow at the base of the snow cover was generally of lower conductivity than the snow at the top) (Fig. 3-20). Thus, the observed curvature during periods of relatively stable air temperature could only be the result of (3), (4), or a combination of both, a point which is discussed in Section 5.4.

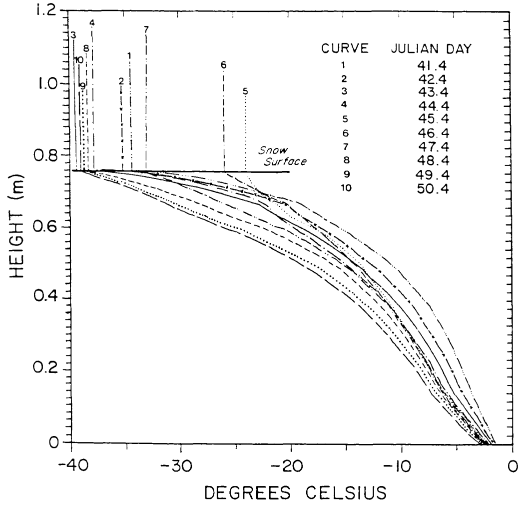

FIGURE 3-29: Typical vertical temperature profiles showing negative curvature. Profiles were measured during a period of relatively stable air temperature.

The curvature was greatest during incoherent events, intermediate during coherent events, and least during isothermal events. This was determined in the following manner: periods of stable air temperature were identified, and a determination of the type of event that was in progress was made from temperature records from the horizontal thermistor strings. Sequential vertical temperature profiles were fit with quadratic curves:

$$(3\text{-}4) \qquad T(z) = a_0 + a_1 z + a_2 z^2 \; (r^2 > 0.99)$$

which were differentiated twice to get the curvature. The range of temperatures along the horizontal strings was measured and plotted against the curvature (Fig. 3-30). The results suggest that the curvature (or the range) can be used, in a general manner, to determine the type of event in progress, particularly for incoherent events. The four fields drawn on Figure 3-30 facilitate the determination. This is a potentially useful result, because it is easier and requires far simpler instrumentation to measure vertical temperature profiles in snow than to measure horizontal profiles using horizontal thermsitor strings.

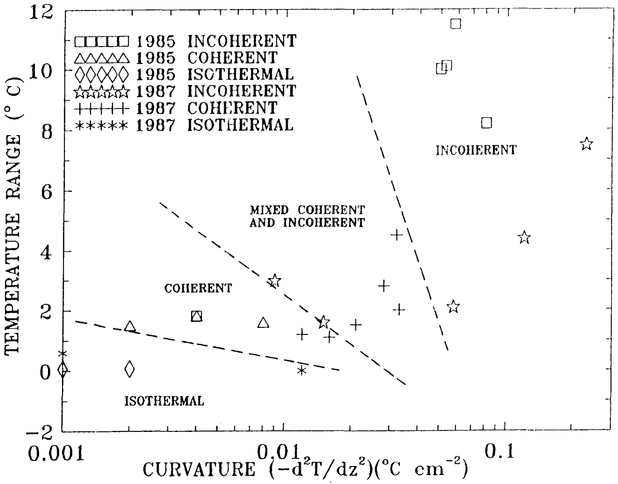

FIGURE 3-30: The type of temperature event as a function of the magnitude of the curvature of the vertical temperature profiles. Curvature is greatest during incoherent events, least during isothermal events. The range of temperature along horizontal strings is also maximum for incoherent events.

4 ANALYSIS AND DISCUSSION OF MASS TRANSPORT IN THE SNOW COVER

4.1 Background

Layer-to-layer mass transport changes the density of snow layers. Within a layer, mass transfer between neighboring grains results in grain growth and shrinkage. (Gubler [1985] referred to the latter case as "interparticle vapor fluxes".) In this chapter, density measurements are used to calculate mass transport between snow layers, and the maximum grain growth rate is estimated.

The growth of large depth hoar crystals (see Section 3.1.3) clearly demonstrates the existence of mass flux gradients between snow grains, but the existence of layer-to-layer mass flux gradients has been disputed. Marbouty (1980) and later Armstrong (1985) found that the mass of a layer of snow remained constant during the growth of faceted depth hoar grains, suggesting that the net layer-to-layer transport was zero. Using this finding, Adams and Brown (1983), Sommerfeld (1983), Colbeck (1983a), Gubler (1985) and Christon et al. (1987) explicitly assumed no net mass transport between layers in their models of dry snow metamorphism. However, Trabant and Benson (1972), working in snow subjected to stronger temperature gradients for a longer period of time than the snow studied by Marbouty (1980) and Armstrong (1985), found significant layer-to-layer mass flux gradients. An important result of the present study is that it confirmed the work of Trabant and Benson (1972). Though the analysis presented below shows that the values determined by Trabant and Benson

(1972) may be slightly high, the essence of their result, i.e., that there are measurable layer-to-layer mass flux gradients in the snow cover, is valid.

The relationship between layer-to-layer mass flux gradients and grain growth is poorly understood. Yosida et al. (1955) postulated that the layer-to-layer flux was the result of innumerable microscopic fluxes between grains in a process they called "hand-to-hand" transfer. In this process, a water molecule accreting to the bottom of an ice grain causes a different water molecule to leave the top of the grain. Repeating the process at successively higher grains results in a net transport of water vapor over a macroscopic distance. The "hand-to-hand" model of vapor transport has become a virtual paradigm in snow science, but it has not been verified nor quantified, and it sheds little light on the relationship between the two scales of mass transport.

Existing theoretical and experimental studies have done little to clarify the relationship between layer-to-layer mass flux gradients and grain growth. Many studies do not explicitly state the scale of the mass transport, and some studies do not even recognize that there are two scales of transport. Yosida et al. (1955), Giddings and LaChapelle (1962), de Quervain (1972), and Trabant and Benson (1972) measured or modeled the layer-to-layer mass flux gradients. Adams and Brown (1983), Sommerfeld (1983), Colbeck (1983a), Gubler (1985) and Christon et al. (1987) developed models of grain growth. Results from both grain- and layer-scale models and measurements have been compared without regard for the differences in scale. There are more models than measurements; there are only 10 measurements of mass transport at either scale (Table 1-1), and, excluding the present study, only a single work (de Quervain, 1958) in which the measurements necessary to determine both scales of transport were made.

4.2 Layer-to-layer Mass Flux Gradients in the Snow

4.2.1 Layer-to-layer Mass Flux Gradients from Comparison of Snow on the Ground and the Tables

The existence of layer-to-layer mass flux gradients in the Fairbanks snow cover, and an estimate of their magnitude, can be determined by comparing the end-of-winter density profiles from the ground and the tables (Fig. 3-1). Over a twenty year period, the snow on the table was denser at the base than the snow on the ground; conversely the top snow layers on the ground were denser than the top layers on the tables. These density contrasts could only be the result of a net upward transfer of vapor from the bottom to the top of the snow cover on the ground. They imply that there was a vapor flux gradient of about $3 \times 10^{-6} kg\ m^{-2} s^{-1} m^{-1}$ (see below).

Consider the average density curves in the center of Figure 3-1. The greater density at the base of the snow on the tables could be partially the result of more rapid compaction on the tables than the ground because of textural differences in the snow, a net upward transfer of mass in the snow on the ground, or a combination of both. But the density contrast at the top of the snow can only be explained by a net upward transfer of mass. At the top, both snow on the tables and snow on the ground had initial densities of approximately 100 kg m^{-3}, but the snow on the ground became more dense than the corresponding snow on the table. With little or no overburden stress at the top of the snow cover, it is unlikely that differing rates of compaction were responsible for the density contrast, which in any case would have resulted in the opposite density contrast, since the snow on the ground was composed almost entirely of depth hoar which was more resistant to compaction than the snow on the tables (see Section 3.1.1). A net accumulation of mass in the upper part of the ground snow cover is the only reasonable mechanism to produce the increased density, as pointed out by Trabant and Benson (1972). The mass must have come from the lower part of the snow cover.

In comparing the density profiles of snow on the ground with those on the tables, it can be seen that the gain in the upper part of the ground snow cover was roughly equal to the loss from the lower part. If in the course of a winter (175 days) the mass (11 kg m^{-2}) moved up and none moved out the top of the snow cover, it would require a layer-to-layer mass flux gradient of 3×10^{-6} kg m^{-2} $s^{-1}m^{-1}$, assuming the top half of the snow cover was a layer 0.25 m thick.

Due to the low air temperatures present throughout the winter in Fairbanks, the snow surface is usually the location of the minimum temperature (see Section 4.2.2). Consequently, it is a site of condensation rather than sublimation. Therefore, assuming little or no water vapor escapes from the snow is reasonable.

4.2.2 Mass Flux Gradients from Layer Density and Thickness Measurements

Layer density and thickness measurements made in 1987 allowed layer-to-layer mass flux gradients to be calculated in a more detailed manner than in the preceding section. For each of the 10 marked layers of snow (see Sections 2.2.1 and 3.1.1) the continuity equation was solved to determine the mass flux gradient. The continuity equation for a compacting layer of snow is (Appendix V):

$$\frac{\rho}{h}\frac{Dh}{Dt} + \frac{D\rho}{Dt} \approx -\frac{\partial J}{\partial z} \qquad \text{(4-1)}$$

where ρ is the bulk density of the snow, t is time, h is the thickness of the snow layer, z is the vertical coordinate, and J is a vertical mass flux, here limited to a vapor flux. Both J and z are positive upward. The minus sign accounts for the fact that when more vapor enters through the bottom than exits through the top of layer (a negative flux gradient) it increases the mass of the layer.

From density and layer thickness data (see Section 3.1.1; Tables 3-2 and 3-3), empirical linear relationships:

$$h(t) = a_1 t + b_1$$

(4-2)

$$\rho(t) = a_2 t + b_2$$

were fitted to the compaction [h(t)] and densification [ρ(t)] curves for each snow layer (Fig. 4-1). The compaction and densification were rapid immediately after the snow was deposited, but this was followed by a longer period of reduced rates, therefore the data for the initial and secondary periods were fit with separate linear relationships. From the functional relationships (Eqns. [4-2]), the derivatives $\frac{D\rho}{Dt}$ and $\frac{Dh}{Dt}$, and ρ and h were evaluated for the midpoint of each line segment, and Equation (4-1) was solved for an early- and late-winter value for the mass flux gradient for each layer (Table 4-1). Uncertainties in thickness (± 3 mm) and density (about ± 5 kg m^{-3}) made the uncertainty in the calculated value of the mass flux gradient relatively large (Appendix V).

The results of the calculations for individual snow layers suggest that the redistribution of mass in the snow took place in a complex and episodic manner (Fig. 4-2 and Table 4-1). For example, the maximum 1987 layer-to-layer mass flux gradient (**26 X 10^{-6}kg $m^{-2}s^{-1}m^{-1}$**), though it only existed for a short interval in a single layer of snow, was an order of magnitude greater than the 20-year average gradient. Several other layers also had flux gradients which were greater than average for at least part of the winter. These brief periods with strong gradients suggest some of the mass transport may have occurred during short, intense bursts of high flux. In several snow layers (i.e., layers 4, 5, 7, and 8) the flux gradient even changed sign during the winter.

Mass flux gradients calculated for individual snow layers (1987) were consistent with the net transfer of mass from the base to the top

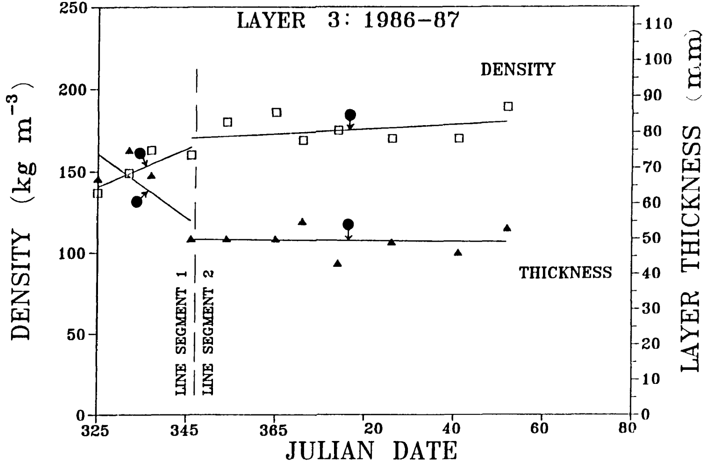

FIGURE 4-1: Line segments fit to a compaction and densification curve for a snow layer on the ground, 1987. The mid-point (●) of each line segment is marked. h, ρ and (Dh/Dt) and (Dρ/Dt) were evaluated at this point.

TABLE 4-1: Calculated values of the layer-to-layer mass flux gradient and flux for snow on the ground, 1987. ρ and h were evaluated on the indicated Julian date. See Appendix V for the methods of calculation. Layers are defined in Table 3-3, Figure 3-5, and Figure 4-2.

LAYER	INTERVAL	Δt	h	ρ	JULIAN DATE	$\partial J/\partial z$	J
		DAYS	mm	kg/m^3		$kg/m^2s\ m$ X 10^{-7}	kg/m^2s X 10^{-7}
1	21NOV-21FEB	91	59	165	14	5.5	2.9
2	21NOV-21FEB	91	19	168	14	18.9	3.3
3	21NOV-15DEC	24	64	155	337	79.4	8.4*
3	15DEC-21FEB	67	49	177	17	35.7	5.0*
4	3DEC-31DEC	28	37	163	350	-70.7	5.8*
4	31DEC-21FEB	52	32	193	33	14.0	5.5
5	15DEC-14JAN	30	26	192	1	263.9	12.6*
5	14JAN-21FEB	38	22	199	33	-17.6	5.1
6	15DEC-21FEB	67	22	194	17	10.5	5.3
7	31DEC-27JAN	27	36	135	13	109.0	9.3*
7	27JAN-21FEB	24	27	180	39	-58.0	3.8*
8	14JAN-27JAN	13	80	157	20	207.2	-7.2*
8	27JAN-21FEB	24	68	181	39	28.6	5.7*
9	14JAN-21FEB	37	50	170	32	67.2	9.1*
10	27JAN-10FEB	14	12	117	34	(?)	(?)

* significant value: not due to uncertainties in calculations
(?) top layer of snow difficult to work with: values unreliable

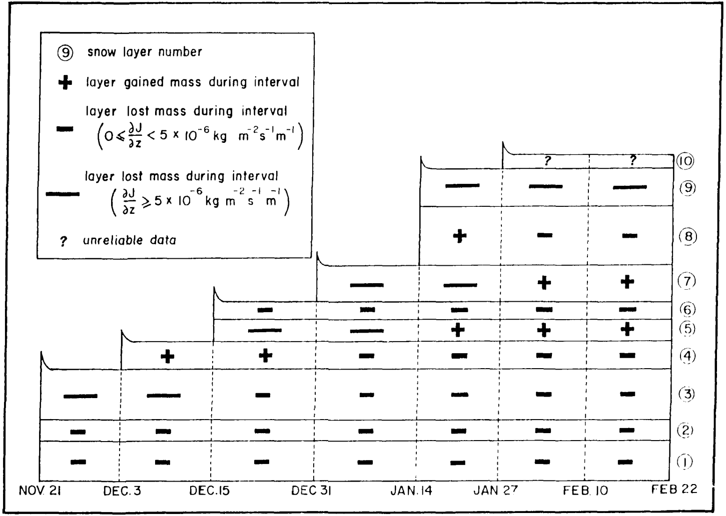

FIGURE 4-2: Layer-to-layer mass flux gradients for the Fairbanks snow cover, 1987 (see also Table 4-1). Minus signs indicate that a layer was losing mass ($\partial J/\partial z \geq 0$); plus signs indicate it was gaining mass ($\partial J/\partial z \leq 0$). See text for discussion.

of the snow implied by the end-of-winter density profiles (see Section 4.2.1). In fact, the end-of-winter profiles on the ground and the tables for 1987 are nearly identical with the 20-year average profiles (Fig. 3-1). Also, the average mass flux gradient for the lower half of the snow cover (layers 1 through 4) calculated from the values for individual layers (a weighted average, where the layer thickness and the time interval were used as weighting factors) was approximately 2 X 10^{-6} kg $m^{-2}s^{-1}m^{-1}$, which is reasonably consistent with the 20-year average.

Two details of the layer-to-layer mass transport for 1987 are not consistent with the pattern suggested by 20 years of end-of-winter density profiles. First, the calculated flux gradients in the lowest two layers of the snow, which were expected to be the source of much of the water vapor transported to the top of the snow, were quite small (Fig. 4-2 and Table 4-1). This may have been because these two layers were deposited in October and had been on the ground nearly 40 days at the time that they were first measured. They were already depth hoar and there might have been significant vapor flux out of the layers before they were measured. Second, zones of mass accumulation ("+" signs on Figure 4-2) were not always near the snow surface. This is puzzling. Nyberg (1938) has shown that minimum temperatures are usually found at the snow-air interface, a fact confirmed by measurements made during this study. Due to this temperature minimum, the interface is a preferential site for the condensation of water vapor from below and above. Condensation from above results in the formation of surface hoar (Colbeck, 1988; Lang et al., 1984), but the data from 1987 indicate that the snow-air interface was not always the locus for accumulation of mass from below.

The average flux gradient for 1987 is about three times greater than the value obtained by Yosida et al. (1955) for snow under much weaker temperature gradients, and about 30% of the value reported by Trabant and Benson (1972) for the Fairbanks snow cover (Table 1-1). Trabant and Benson (1972) calculated the mass flux gradient without measuring changes in the thicknesses of snow layers. Instead they used the densification rate of snow perched on the tables to evaluate the rate at which snow layers on the ground compacted. It was assumed that a snow layer on the table was subject only to compaction, because there was no strong temperature gradient to induce a vapor flux. In that case, in Equation (4-1) $(\frac{\partial J}{\partial z})_{TABLE}$ was zero, and $(\frac{D\rho}{Dt})_{TABLE}$ equaled $(\frac{\rho}{h}\frac{Dh}{Dt})_{TABLE}$. It was further assumed that the compaction rate $(\frac{Dh}{Dt})$ and the ratio, $\frac{\rho}{h}$, were the same on the tables as on the ground, so that $(\frac{\rho}{h}\frac{Dh}{Dt})_{TABLE}$ equaled $(\frac{\rho}{h}\frac{Dh}{Dt})_{GROUND}$. The flux gradient on the ground could then be evaluated by subtracting the densification rate on the table from the densification rate on the ground:

$$(4\text{-}3) \qquad (\frac{D\rho}{Dt})_{GROUND} - (\frac{D\rho}{Dt})_{TABLE} \approx -\frac{\partial J}{\partial z}.$$

Unfortunately, the critical assumption behind this method is incorrect. Figure 3-4 shows that for two stratigraphically equivalent layers, the layer on the table compacted more quickly than the layer on the ground, particularly immediately after deposition. Later in the winter their compaction rates were nearly equal. The difference in compaction rates makes the use of Equation (4-3) appropriate only long after the snow has been deposited. The compaction rates differed because the snow deposited on the tables did not metamorphose into depth hoar, which resists vertical compaction (see Section 3.1.1).

The measurements of Trabant and Benson (1972) were reinterpreted making use of the fact that the measured compaction rates on the tables and the ground were nearly equal several weeks after the snow had been deposited. The available data only allowed recalculation for the bottom 0.13 m of the snow. Here, the recalculated value of the mass flux gradient averaged 5×10^{-6} kg $m^{-2} s^{-1} m^{-1}$ for four winters, which is about twice the average value calculated in the present study for 1987. It is likely that the recalculated value is high because even during the period of reduced compaction and densification rates, the snow on the tables still compacted slightly faster than the snow on the ground (Fig. 3-4).

It is possible to use the vapor flux measured at the snow-soil interface (0.26×10^{-6} kg $m^{-2} s^{-1}$; see Table 3-1) as the starting point to integrate the flux gradients layer-by-layer up through the snow cover in order to estimate the flux out of the top of each layer (Table 4-1 and Appendix V). The maximum flux computed in this manner was 1.26×10^{-6} kg $m^{-2} s^{-1}$, although for most of the layers most of the time the flux was rarely more than twice the flux at the snow-soil interface.

4.3 Estimates of Grain Growth Rates

The grain size and number data presented in Section 3.1.2 show that only a few selected snow grains (about 1 out of 10) grow during the development of depth hoar. The majority of grains shrink to supply mass to the growing grains. Layer-to-layer mass flux gradients can also supply mass to the growing grains. Grains must occupy favorable positions and be oriented favorably in order to grow. Even then, as conditions change within the snow cover, growing grains may undergo periods of shrinkage, or grow on one side and shrink on another (i.e., the erosion textures seen on the upper surfaces of crystals in Metamorphic Layer 4; see Frontispiece). Compaction of the snow can bring two grains into contact so that they grow together into a single grain. These physical processes make modeling grain growth difficult. In order to be consistent with observed growth behavior, a realistic grain growth model must be a function of 1) grain size, 2) grain size distribution, 3) grain position with respect to adjacent grains, 4) c-axis orientation, 5) temperature, and 6) over-burden stress. At present, no grain growth models are appropriate for use with the grain size distribution data collected in this study.

The data collected in this study were sufficient to make an order of magnitude estimate of the maximum growth rate of depth hoar. The masses of 20 of the largest depth hoar grains which grew by April 8, 1986 were estimated from photographs. They varied between 1 and 2 X 10^{-4} kg, with a maximum value of 5 X 10^{-4} kg. These grains began to grow in December at which time their mass was several orders of magnitude less than their final mass and can be considered essentially zero. Over 130 days, their average growth rate was 1×10^{-11} kg s^{-1} (0.9 mg day^{-1}), with a maximum rate of 5×10^{-11} kg s^{-1} (4.3 mg day^{-1}).

5 ANALYSIS AND DISCUSSION OF HEAT TRANSPORT IN THE SNOW COVER

In this chapter, the temperature data from Chapter 3 are used to show that air convected through the snow. The 1985 data showed convection occurred (Johnson et al., 1987), though the analysis was limited to a single, ten-day event early in the winter. Here, the data are used to show that convection occurred during all three winters of the study. A description of the convective circulation is developed from the patterns observed in the horizontal temperature field and by analogy with laboratory experiments. The air flow velocity during periods of convection is estimated using one dimensional heat and mass transport equations. Using this velocity estimate and measured values of the layer-to-layer mass flux gradient (Chapter 4), the importance of convection in moving heat and mass is discussed.

5.1 The Evidence for Convection in the Snow Cover

There is evidence for time-dependent convection in the Fairbanks snow cover. A quasi-stable convective flow, indicated by coherent temperature signals (see Section 3.2.3 and Fig. 3-22b), set up in early winter and evolved slowly as the basal boundary conditions changed. Convection was present during 80% of the winters of 1986 and 1987. It was absent only during the few brief thaws when vertical temperature gradients in the snow approached zero (Figs. 3-25 and 3-26). The flow circulation pattern was controlled by horizontal differences in temperature along the snow-soil interface. Diffuse, plume-like updrafts developed above regions of higher temperature. As winter progressed and the soil moisture froze, temperature contrasts at the snow-soil interface changed (Fig. 3-28), causing the slow evolution of the convective circulation pattern (Fig. 3-25).

During brief periods (a total of about 15% of the winter), external forcing mechanisms such as high wind or rapid changes in air temperature perturbed the convection (Figs. 3-23 and 3-24). During these periods, the flow circulation pattern remained relatively unchanged (Fig. 3-25), but the strength of the convection (i.e., the flow speed) increased. The perturbations in the convection resulted in incoherent temperature signals (see Section 3.2.3 and Fig. 3-22c). These showed greater horizontal temperature gradients than the coherent temperature signals characteristic of the quasi-stable convective flow (16 K m^{-1} vs. 10 K m^{-1}). Crossing of thermistor traces and the simultaneous warming and cooling of different thermistors on a single horizontal string, distinctive features of the incoherent temperature signals (Fig. 3-22c), were the result of thermal disequilibrium in the snow as it adjusted to the new flow conditions. (These features may also have been, in part, the result of natural spatial variations in the forcing mechanisms [i.e., small-scale spatial variations in wind speed]). When external forcing ceased, quasi-stable convective flow resumed, and thermal equilibrium was re-established, resulting in a gradual change from incoherent to coherent temperature signals. In contrast, the onset of incoherent events was abrupt (Figs. 3-23 and 3-24).

5.1.1 Incoherent Temperature Signals as Evidence of Convection

The key feature of the incoherent events was that warming and cooling occurred simultaneously at different points on a horizontal thermistor string (Figs. 3-22c, 3-23). The only plausible explanation for this behavior was air movement through the snow. The thermistors which warmed were located in warm updrafts, while those which cooled were located in cold downdrafts.

The close association between incoherent events and periods of strong wind (Table 3-5) suggests that forced convection occurred in some cases. Clarke et al.(1987) and Colbeck (in prep.) have concluded that wind-caused pressure fluctuations can pump air through the snow.

The cause of the incoherent events which did not correlate with high wind is not known. One speculation is that they were caused by other forcing mechanisms. Rapid changes in air temperature is one possible mechanism (see Fig. 3-24; event B); another is pressure fluctuations caused by gravity waves. Wilson and Nichparenko (1967) and Wilson and Fahl (1969) measured gravity waves in Fairbanks which produced atmospheric pressure fluctuations of 10 to 30 μbar. Even though pressure measurements were not made during our investigation, these studies indicate that periods of constant pressure do not exist.

5.1.2 Coherent Temperature Signals as Evidence of Convection

The possibility that conditions other than convection could have produced coherent temperature signals was examined. Several points in the following analysis also apply to incoherent temperature signals.

In the absence of convection, coherent temperature signals could have arisen from:

1) improper thermistor calibration,
2) horizontal variation of the thermal conductivity,
3) variations in thermistor heights,
4) variations in the thickness of the snow cover, and
5) relatively warm and cold spots at the snow-soil interface.

(1) can be dismissed, because the thermistors were accurate to ± 0.03°C, 500 times more sensitive than the observed temperature range (see Section 2.3.2 and Appendix II).

(2) is unlikely because there is no evidence to suggest horizontal variations in the thermal conductivity of the snow. Also, considering that the total measured range in thermal conductivity was quite small (Appendix III), any horizontal variations in the thermal conductivity, if they had existed, would have been insignificant.

(3) and (4) were known to be potential problems at the start of the study, so considerable effort was taken to avoid them. The height of each thermistor was measured when it was installed and when it was excavated at the end of the winter (see Section 2.3.2 and Fig. 3-27). The observed variations in height were compared with the variations in

height that would have been required to produce the temperature deviations by conduction. From the vertical thermistor strings, the temperature gradient at the height of the horizontal strings was determined. The range in temperature along the horizontal strings was measured, and the vertical misplacement which would have been required to produce the full range was computed. For example, if the vertical gradient was 100 K m^{-1} at the height of the thermistors, then a horizontal temperature range of 10° would require a vertical misplacement of 0.1 m. Johnson et al. (1987) analyzed the 1985 data in this manner; Table 5-1 lists the results for 1986 and 1987. For all three winters, the measured vertical misplacements were 1/3 to 1/10 the amount necessary to produce the observed temperature range. Not only were the misplacements too small, but also the relative thermistor misplacements would have had to change with time, which is physically unreasonable. Therefore, it is unlikely that the temperature deviations were the result of thermistors at different heights.

Since the snow-soil interface was level (see Section 2.3.2), local variations in the snow depth (4) could only have been the result of surface drifting. Surface drifts rarely exceeded 0.01 m (Fig. 3-27a). The effects of surface drifts were examined using a steady-state, two-dimensional finite difference conduction model. Results from the model showed that drifts in excess of 0.06 m would have been required to produce the observed range of horizontal temperatures, and as more snow accumulated during the winter, the height of the drifts would have had to increase, making drifting an unlikely cause of the

TABLE 5-1: Observed vs. required vertical misplacement of thermistors.

JULIAN DATE	$\partial T/\partial z$ @ THERMISTOR [$K\ m^{-1}$]	OBSERVED TEMPERATURE RANGE [K]	REQUIRED VERTICAL MISPLACEMENT [m]
		1986	
Ave. end-of-winter thermistor ht.:		0.119 m	
Std. deviation:		± 0.010 m	
Number of thermistors:		145	
341	-41.9	5.0	.060
349	-45.0	5.0	.056
2	-37.2	3.0	.040
13	-34.4	6.0	.087
41	-13.8	2.5	.091
		1987	
Ave. end-of-winter thermistor ht.:		0.110 m	
Std. deviation:		± 0.007 m	
Number of thermistors:		152	
330	-187.0	8.0	.021
345	1.0	0.1	.063
358	-106.3	5.0	.024
8	-19.8	1.0	.025
36	-41.6	2.5	.030

observed temperature deviations.

The poor correlation between relative warm and cold zones in the snow with a) the thermistor heights and b) the snow depth (see Section 3.2.3) reinforces that these two factors were not the cause of the coherent temperature records. If differences in the heights of the thermistors or the amount of snow above the thermistors had affected their temperatures, the correlation coefficients would have been high, but instead they never exceeded 0.36.

(5) The significant spatial variations in the temperature of the snow-soil interface (Fig. 3-28) could not have produced the observed temperature signals by conduction alone. However, they may have had a profound effect on the stability of the air in the snow and the initiation of convection, as discussed in Section 5.2.2. The maximum range in temperature at the interface was 4°C, while the maximum range in temperature on the horizontal strings 0.1 to 0.2 m above the interface was often greater than 5°C, occasionally as high as 16°C. Clearly, the temperature range at the thermistor strings could not have been the result of heat conducted from the interface. A second steady-state, finite difference model was used to investigate the effect of a warm spot at the snow-soil interface. For a 0.6 m deep snow cover with a basal boundary containing a 1 m wide spot elevated 3°C above the normal interface temperature, the maximum temperature contrast that would be observed 0.1 m above the interface would be 1.5°C. For smaller warm zones the temperature contrast would be even less. The observed horizontal variations in temperature were almost

an order of magnitude greater than 1.5°C, and could not have been produced by conduction.

The argument could be made that combinations of mechanisms (1) through (5) produced the observed temperature signals by conduction, but this is unlikely. It would have required the alignment of the crests of snow drifts above just the thermistors which had been displaced downward, and directly above warm spots at the snow-soil interface. If snow drifts, warm spots, and displaced thermistors occurred randomly, the chances of this sort of alignment would be low. One is forced to conclude that the coherent temperature signals were the result of air convecting through the snow.

The rapidity with which the snow responded to changes in air temperature is also strong evidence for convection during both coherent and incoherent events. The snow at the depth of the horizontal thermistor strings responded to changes in air temperature ten times faster than it should have by heat conduction alone. In the absence of convection, the bulk thermal diffusivity ($\alpha = k/\rho c$) was approximately $1 \times 10^{-7} m^2 s^{-1}$. Because there was little variation in either k or ρ with height (Figs. 3-2, 3-20), the bulk thermal diffusivity determined for the entire snow cover was approximately the same as for each layer. The time necessary for a change in air temperature to propagate to a depth, L, in the snow is approximately $\tau = (L^2/\alpha)$ (Turcotte and Schubert, 1982). Using measured values of the snow depth (h) and horizontal thermistor string height (z_t) to calculate L ($L = h - z_t$), the conductive response time of a number of rapid changes in air temperature have been calculated and compared to the actual response time (Table 5-2). In all cases, during coherent or incoherent events the response was one to two orders of magnitude faster than predicted. In order to respond this fast to changes in temperature, the air in the snow must have been in motion. The rapidity with which the snow responds to changes in air temperature has been noted by other authors (Bey, 1951; Benson, 1962; Fay, 1973), which suggests that convection occurs elsewhere.

In laboratory experiments, the presence of convection is usually indicated by an increase in heat transfer. This is often shown using the Nussult number (Nu), which is the ratio of total heat transfer to conductive heat transfer; it can be expressed as:

$$\text{(5-1)} \qquad Nu = \frac{k_{eff}}{k}$$

In the absence of convection, Nu equals 1; when there is convective heat transfer, it is greater than 1. In the present study, the Nussult number was calculated by comparing the effective bulk thermal conductivity of snow measured by heat flow meters (Fig. 3-18) to the average effective thermal conductivity measured by needle probe (see Section 3.2.2). The effective bulk conductivity ranged from 0.1 to 0.2 W $m^{-1}K^{-1}$; the average thermal conductivity ranged from 0.04 to 0.10 W $m^{-1}K^{-1}$. Comparing these values suggests that the Nussult number was about 2, which is consistent with the conclusion reached above that convection was prevalent in the Fairbanks snow cover.

TABLE 5-2: Response time of the snow to changes in air temperature, 1985-1987. Predicted lag time of the snow has been calculated assuming conduction as the only mode of heat transport using the equation in the text (Section 5.1.2). The ratio is the predicted lag time divided by the observed lag time. α is bulk thermal diffusivity of the snow.

JULIAN DATE	TYPE OF EVENT	# OF OBS.	SNOW DEPTH [m]	THERM. HT. [m]	PREDICTED LAG [s]	OBSERVED LAG [s]	RATIO
1985							
344-351	INCO.	6	0.24	0.18*	36,000	7,200	5
351-360	CO.	2	0.84	0.18*	4,400,000	132,500	33
42--52	INCO.	1	0.73	0.18*	3,000,000	59,500	51
74--84	CO(?).	1	0.75	0.17*	3,400,000	125,000	27
95-105	ISO(?).	2	0.70	0.16*	2,900,000	228,000	13
1986							
345-350	CO(?).	7	0.26	0.18*	64,000	5,400	12
350-355	CO.	7	0.26	0.18*	64,000	5,400	12
355-360	CO.& INCO.	5	0.25	0.17*	64,000	3,900	16
360-365	CO.	8	0.25	0.17*	64,000	6,900	9
365---5	CO.& INCO.	3	0.25	0.17*	64,000	7,500	9
55--65	CO.& INCO.	13	0.28	0.16*	144,000	14,800	10
1987							
345-350	CO.	7	0.23	0.14	81,000	3,200	25
350-355	CO.	7	0.28	0.14	196,000	6,000	32
355-360	CO.	11	0.22	0.13	85,000	8,900	10
360-365	CO.	9	0.24	0.12	132,200	6,600	20
365---5	INCO.	2	0.24	0.12	139,000	4,900	28
25--30	CO.& INCO.	2	0.42	0.12	900,000	32,300	28

CO.= Coherent Event INCO.= Incoherent Event
ISO.= Isothermal Event (?)= Uncertainty * = Estimated Value
Predicted Lag Calculated using $\alpha = 1 \times 10^{-7} m^2 s^{-1}$

5.2 The Nature of the Convective Circulation

One of the striking characteristics of the temperature field in the snow is that relative warm and cold zones developed and remained nearly stationary throughout the winter (Figs. 3-25, 3-26). During the winter these zones evolved slowly, with their relative temperatures changing through time. Their evolution was independent of the snowfall history and snow settlement (Fig. 3-5), suggesting that the zones were caused by a mechanism which was not in the snow. Also, where it was possible to compare the location of the zones with the temperature at the snow-soil interface, it was found that warm zones coincided with the higher temperatures. These facts suggest that the warm and cold zones in the snow were caused by variations in temperature at the snow-soil interface due to unevenly distributed surface moisture (see Section 3.2.3). Since the variation in temperature at the interface could not account for the warm and cold zones by conduction (see Section 5.1.1), we suggest that it generated a plume-like convective circulation system. Diffuse plumes of warm, rising air were centered over the warmer areas of the interface, while cold downdrafts formed elsewhere. As shown below, this hypothesis is consistent with the temperature observations and with the results of laboratory studies of convective plumes.

5.2.1 The Convective Circulation

The temperature fields shown in Figure 3-25 do not indicate the periodic spatial repetition that would result from cellular or roll convection (Rayleigh, 1916; Combarnous and Bories, 1975; Powers et al., 1985a). In 1985, the spatial resolution of the thermistor grid was insufficient to rule out cellular circulation, but the temperature patterns developed in the next two winters show no repetition, even on the scale of expected cell size (cells of about 0.5 m). It is likely that conditions in the snow in 1985 were similar.

The temperature patterns, instead, suggest that there were at most three convective plumes, and usually only one or part of one convective plume intersected by the thermistor grid. The plumes were diffuse, with diameters of about a meter or less (Fig. 3-25). Downdrafts had slightly smaller diameters of about 0.3 m. Plume updrafts were between 2° and 16°C warmer than nearby downdrafts, and both up- and downdrafts were relatively fixed in space throughout the winter (see correlation analysis, Fig. 3-26). Plumes were present during both coherent and incoherent events, but were presumably stronger during incoherent events, because the temperature contrasts between up- and down-drafts were greater. In 1987, when snow-soil interface temperatures were measured at the south end of the horizontal thermistor array (Fig. 3-27), plume updrafts were linked to the higher temperatures. Similar coupling between plumes and warm spots at the snow-soil interface is assumed to have existed elsewhere.

Laboratory experiments support the hypothesis that convective plumes developed above warm spots at the base of the snow. Experimental and theoretical studies in porous media by Elder (1967) and El-Khatib and Prasad (1986) indicate that a circulation pattern similar to the one shown in Figure 5-1 would develop above a warm spot. Plumes or updrafts would be narrow and of higher velocity than downdrafts, so in horizontal cross-section, downward flow would occupy a larger area than upward flow. Recharge and discharge would occur through the permeable upper surface of the medium, possibly resulting in a net gain or loss of mass and a recirculation of about 30% of the fluid. Flow lines would be predominantly vertical except along the base where they would be horizontal.

Laboratory and theoretical studies also support the hypothesis that the convective circulation system which developed in the snow cover was time-dependent. For example, Schubert and Straus (1982) have shown that varying the boundary conditions of a porous medium, such as occurs during rapid fluctuations in air temperature, can produce transient convection states with distinct oscillatory regimes. These regimes can change from polyhedral cellular convection to a state in which circulation cells appear and disappear rapidly (Combarnous and Bories, 1975). Even with constant conditions, long periods are required to achieve steady state, during which transitory circulation patterns come and go (Elder, 1967). Even small perturbations in the snow may have had large effects on the circulation system. Meyer et al. (1987) showed that the circulation patterns which formed in a Benard cell were different each time the

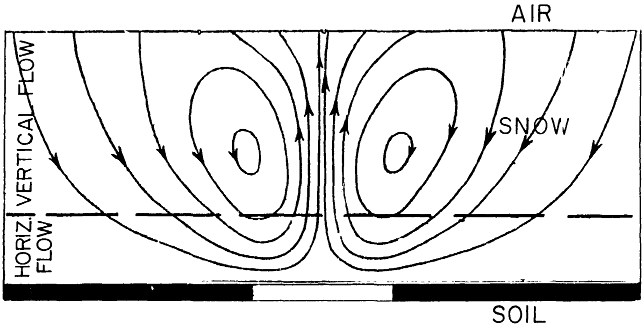

FIGURE 5-1: Hypothetical air flow circulation over a warm spot (white), based on Elder (1967). Flow lines are predominately vertical except at the base of the snow. 30% of the air recirculates; there is discharge from the plume above the warm spot.

experiment was run, pointing to the importance of stochastic effects on the flow pattern.

5.2.2 The Onset of Convection

Convection occurs when the buoyant forces exceed the viscous resistance of a fluid. The ratio of buoyant to viscous forces is usually presented as a dimensionless ratio, the Rayleigh number (Ra), which can be written:

$$(5\text{-}2) \qquad Ra = \frac{g\ \beta\ (\rho c)_f\ \ \Delta T\ h\ \kappa_i}{\nu\ k_m}$$

where g equals the acceleration of gravity, $\beta = \frac{1}{V}\left(\frac{\partial (V)}{\partial T}\right)_p$ is the isobaric coefficient of thermal expansion, $(\rho c)_f$ is the volumetric heat capacity of the fluid (moist air), ΔT is the temperature difference across the snow layer, h is the layer thickness, κ_i is the coefficient of air permeability, ν is the kinematic viscosity and k_m is the thermal conductivity of the porous medium (Lapwood, 1948; Katto and Masuoka, 1967; Akitaya, 1974; Combarnous and Bories, 1975; Turcotte and Schubert, 1982). This formulation includes the following approximations:

1) the Boussinesq approximation: variations in the fluid density are negligible in all but the buoyancy term,
2) the fluid is incompressible,
3) inertial forces in the fluid are negligible,

4) second order terms are dropped (linear stability analysis),

5) the system is in steady-state,

6) the boundaries are horizontal and isothermal, and

7) there is no latent heat transfer (a source term).

Several authors have determined the critical Rayleigh number (Ra_{cr}) (i.e., the Rayleigh number at which convection begins) for conditions thought to be appropriate for snow. Lapwood (1948) and Nield (1968) determined that Ra_{cr} equaled about 27 for a horizontal porous layer heated from below with isothermal, impermeable upper and lower boundaries. Ribando (1977) found Ra_{cr} equaled 40 for a constant temperature, permeable top, and a constant heat flux impermeable base.

Based on a critical Rayleigh number between 27 and 40, it has been concluded that convection is unlikely in most natural snow covers (Palm and Tveitereid, 1979; Powers et al., 1985a,b; see Section 1.1). However, the Rayleigh number given by Equation (5-2) cannot be appropriate because it never exceeded 27 during this study, despite the fact that air was convecting in the snow almost continuously (Fig. 5-2). Convection was observed for Rayleigh numbers as low as 5 (for example on March 1, 1985).

One reason that the Rayleigh number in Equation (5-2) fails to predict the onset of convection is that approximations (5), (6), and (7) cannot be applied to a natural snow cover. In particular, the presence of spatial variations in temperature at the snow-soil interface undoubtedly lowered the critical Rayleigh number and may produce a situation in which convection will always occur (B. Travis, personal communication, 1988). Without a correct formulation of the

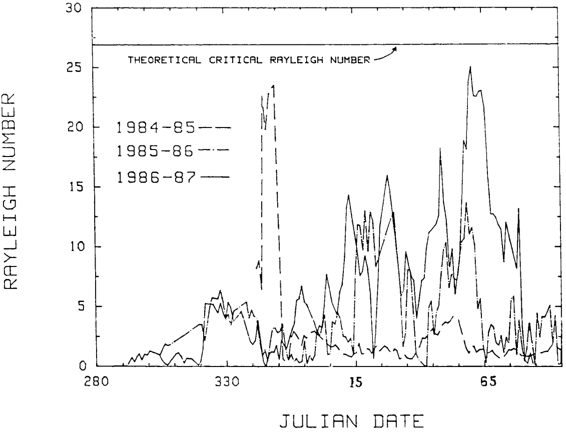

FIGURE 5-2: The Rayleigh number, 1985-1987. The Rayleigh number was calculated according to Equation (5-2) using measured values for the Fairbanks snow cover. The theoretical critical Rayleigh number for the onset of convection thought to be applicable for snow is 27.

Rayleigh number for natural snow covers as a guide, it is difficult to predict convection in snow.

5.3 The Convective Flow Velocity

In this section we estimate the convective flow velocity. An order of magnitude estimate is made using measured values of the mass flux gradient. A second estimate is made by combining heat and mass transport equations and solving for the velocity.

5.3.1 Mathematical Framework

For a small volume of snow, assuming the air and ice are in local thermal equilibrium (see Section 5.3.3), the heat transport is given by:

$$(5\text{-}3) \qquad \nabla\cdot(k_{dry}\,\nabla T) + L_s \left.\frac{\partial\rho}{\partial t}\right]_v - \phi(\rho c)_f \vec{v}\cdot(\nabla T) = (\rho c)_m \frac{\partial T}{\partial t}$$

where T is temperature, L_s is latent heat of sublimation (Table 5-3), $\left.\frac{\partial\rho}{\partial t}\right]_v$ is the densification rate of the snow due to the vapor flux, ϕ is the porosity, $\vec{v}$ is the air flow velocity through the pores of the snow, $(\rho c)_f$ is the volumetric heat capacity of air, and $(\rho c)_m$ is the volumetric heat capacity of the bulk snow ($[\rho c]_s$ is the volumetric heat capacity of ice) (Table 5-3). The appropriate thermal conductivity, k_{dry}, is that of snow measured in a way which precludes convection, and at temperatures low enough that there is no latent

TABLE 5-3: Physical parameters of the Fairbanks snow cover.

Thermal Conductivity (k)

New snow	$k = 5 \times 10^{-2}$ W m^{-1} K^{-1}	for $\rho \approx 50$ kg m^{-3}
Depth hoar	$k = 4 \times 10^{-2}$ " "	"Ice
"Dry" snow	$k_{dry} = 2.5 \times 10^{-2}$ " "	"Snow
Air	$k_{air} = 2.1 \times 10^{-2}$ " " "	@ -10°C
Ice	$k_{ice} = 2.3 \times 10^{0}$ " " "	@ -10°C

Latent Heat of Sublimation (L_s)

$$L_s = 2.834 \times 10^{6} \text{ J kg}^{-1}$$

Diffusion Coefficient of Water Vapor in Air (D_o)

$$D_o = 2.1 \times 10^{-5} \text{ m}^2 \text{ s}^{-1}$$

Specific Heat (c)

Air	$c = 1.01 \times 10^{3}$ J kg^{-1} K^{-1}	for 0 to -40°C
Ice	$c = 2.09$ " " " " "	@ 0°C
	$c = 1.97$ " " " " "	@ -20°C
	$c = 1.80$ " " " " "	@ -40°C

Density (ρ)

Air	$\rho_{air} = 1.5$ kg m^{-3}	@ -10°C
Ice	$\rho_{ice} = 916.0$ " "	
Depth Hoar	$\rho \approx 200$ " "	

Porosity (ϕ)

$$\phi = \{1 - \frac{\rho}{\rho_{ice}}\}$$

Volumetric Heat Capacity (ρc)

Air
$(\rho c)_f = 1.507 \times 10^{3}$ J m^{-3} K^{-1}
$(\rho c)_s = 1.805 \times 10^{6}$ " " "
$(\rho c)_m = (\rho c)_s (1-\phi) + (\rho c)_f \phi$: where ϕ = porosity

Thermal Diffusivity (α)

Air	$\alpha_{air} = 1.556 \times 10^{-5}$ m^2 s^{-1}	
Ice	$\alpha_{ice} = 1.280 \times 10^{-6}$ " "	
Snow	$\alpha_{snow} \approx 1 \times 10^{-7}$ " "	for ρ=200 kg m^{-3} @ -10

Saturation Water Vapor Density (empirical fit to data)

$$\rho_v(T) = A\, e^{B\,T}$$ where T is in °C; ρ_v in kg m^{-3}

A = 0.005789 B = 0.09658

Dimensionless Quantities

$$\frac{(\rho c)_f}{(\rho c)_s} \approx .001 \qquad \frac{(\rho c)_m}{(\rho c)_f} \approx 280 \qquad \frac{\alpha_{air}}{\alpha_{snow}} \approx 156$$

Dimensionless Ratio of Sensible to Latent Heat Capacity

$$\gamma = \frac{(\rho c)_f}{L_s \left(\frac{\partial \rho_v}{\partial T}\right)}$$

TEMPERATURE	γ
-1°C	1.5
-20°C	6.5
-40°C	48.2

heat transfer. For depth hoar, it equals 0.025 W $m^{-1}K^{-1}$.

The mass transport for a snow layer which is not compacting is given by:

$$(5\text{-}4) \qquad \left.\frac{\partial \rho}{\partial t}\right]_v = \nabla\cdot[\phi F D_o \nabla\rho_v] - \frac{\partial(\rho_v \phi)}{\partial t} - \nabla\cdot[\phi\ \rho_v\ \vec{v}]$$

where ρ is the bulk density of the snow, ρ_v is the water vapor density, F is an enhancement factor discussed in Section **5.3.3**, and D_o is the diffusion coefficient of water vapor in air (Table 5-3).

Due to limitations in the data, Equations (5-3) and (5-4) must be simplified to one dimension to determine the velocity. This can be justified because in many areas in the snow flow lines and temperature gradients are primarily vertical (Fig. 5-1). In one dimension, the equations become:

$$(5\text{-}5) \; k_{dry} \frac{\partial^2 T}{\partial z^2} + L_s \left.\frac{\partial \rho}{\partial t}\right]_v - \phi(\rho c)_f v_z \frac{\partial T}{\partial z} = [\phi(\rho c)_f + (1-\phi)(\rho c)_s] \frac{\partial T}{\partial t}$$

$$(5\text{-}6) \qquad \left.\frac{\partial \rho}{\partial t}\right]_v = \phi\ D_o \frac{\partial \left(F \frac{\partial \rho_v}{\partial z}\right)}{\partial z} - \phi \frac{\partial \rho_v}{\partial t} - \phi \frac{\partial(\rho_v v_z)}{\partial z}$$

where $(\rho c)_m$ of Equation (5-3) is replaced by $[\phi(\rho c)_f + (1-\phi)(\rho c)_s]$ in Equation (5-5), and the following assumptions and approximations have been made:

1) k_{dry} is assumed to be constant in the volume for which the equations are written,
2) there are no horizontal temperature gradients,
3) horizontal components of the velocity field are zero,
4) horizontal components of the diffusive flux are zero, and
5) porosity (ϕ) is constant within the volume and over the time intervals for which the equations are written.

5.3.2 The Convective Velocity Using the Measured Mass Flux Gradient

It is possible to make an order of magnitude estimate of the velocity using the measured layer-to-layer mass flux gradients. This can done by integrating both sides of Equation (5-5) over a time interval, t_o to t_1, and dividing by the elapsed time, Δt, to get time-averages of each term. The densification rate of the snow, $(\frac{\partial \rho}{\partial t}|_v)$ in Equation (5-5), equals the negative mass flux gradient $(-\frac{\partial J}{\partial z})$ (Appendix V). The time average of this term $(-\overline{\frac{\partial J}{\partial z}})$ was measured and is listed in Table 4-1. The other integrations must be over the same time interval. The integral of the right-hand side of Equation (5-5):

$$(5\text{-}7) \qquad \frac{1}{\Delta t} \int_{t_o}^{t_1} \{[\phi(\rho c)_f+(1-\phi)(\rho c)_s] \frac{\partial T}{\partial t}\} \, dt$$

is approximately zero. Assuming ϕ, k_{dry}, and $(\rho c)_f$ are constants, the integral of the first term on the left side of (5-5):

$$\overline{\frac{\partial^2 T}{\partial z^2}} = \frac{1}{\Delta t} \int_{t_o}^{t_1} \{\frac{\partial^2 T}{\partial z^2}\} \, dt \qquad (5\text{-}8)$$

can be evaluated. Unfortunately, the integral of the third term:

$$\overline{\{v_z \frac{\partial T}{\partial z}\}} = \frac{1}{\Delta t} \int_{t_o}^{t_1} \{v_z \frac{\partial T}{\partial z}\} \, dt \qquad (5\text{-}9)$$

cannot be evaluated directly because both v_z and $\frac{\partial T}{\partial z}$ are functions of t. If they were independent functions, Equation (5-9) would equal $\frac{1}{\Delta t} \int v_z \, dt \int \frac{\partial T}{\partial z} \, dt$, and Equation (5-5) could be solved for $\overline{v_z}$:

$$\overline{v_z} = \frac{1}{\phi (\rho c)_f \overline{\frac{\partial T}{\partial z}}} \{k_{dry} \overline{\frac{\partial^2 T}{\partial z^2}} - L_s \overline{\frac{\partial J}{\partial z}}\}. \qquad (5\text{-}10)$$

where $\overline{v_z}$ equals $\frac{1}{\Delta t} \int v_z \, dt$ and $\overline{\frac{\partial T}{\partial z}}$ equals $\frac{1}{\Delta t} \int \frac{\partial T}{\partial z} \, dt$. But because they are not independent, an error is introduced which is approximately equal to the product of the standard deviations of v_z and $\frac{\partial T}{\partial z}$ multiplied by the correlation coefficient between the two functions (Davis, 1973). Assuming the worst case, with the correlation coefficient equal to 1, the approximation results in no more than a 30% error in v_z.

The estimated velocity was 0.2 mm s^{-1} for the average measured values of the mass flux gradient, temperature gradient, and curvature.

The maximum velocity was 1.3 mm s^{-1} (Table 5-4). Since this value is a time average, short periods of higher velocity flow are not precluded.

5.3.3 The Convective Velocity Assuming the Air in the Snow is Saturated

The air flow velocity can also be estimated using a method which does not require measured values of the layer-to-layer mass flux gradient. However, it does depend on the common assumption (see Bader et al., 1939; Giddings and LaChapelle, 1962; de Quervain, 1972; Colbeck, 1982a) that the air in the snow is saturated with respect to ice. This is a reasonable assumption; if it was not, crystal growth and decay rates in the snow would be high. Instead, they are very low. For example the maximum measured growth rate in the Fairbanks snow was 5 X 10^{-11} kg s^{-1} (see Section **4.3**), compared to rates two orders of magnitude higher (as high as 3 X 10^{-9} kg s^{-1}) observed for crystals growing in a diffusion chamber where supersaturation was maintained (Lamb and Hobbs, 1971).

The assumption might not be valid if air was flowing through the snow at a sufficiently high speed. However, a simple model presented below shows that the air would have to flow several times faster than the estimate in Section 5.3.2. Consider a block of ice in which there is a vertical, cylindrical hole of radius R through which air is flowing at a velocity, v_z, and further, that there is a linear vertical temperature gradient ($\frac{\partial T}{\partial z}$) across the block. At any height,

TABLE 5-4: Average flow velocity calculated from measured layer-to-layer flux gradients. The curvature varied as a function of z, but its average value was -100 to -500 K m^{-2}.

INTERVAL	HT	$\overline{\partial T/\partial z}$	$\partial^2 T/\partial z^2$	$\overline{\partial J/\partial z}$	$\overline{v_z}$
	(m)	(K m^{-1})	(K m^{-2})	(kg $m^{-2}s^{-1}m^{-1}$)	(mm s^{-1})
21NOV-21FEB	0-0.06	-35	-1000	6 X 10^{-7}	0.51
"	"	"	- 500	"	0.27
"	"	"	-100	"	0.08
21NOV-21FEB	0.06-.12	-50	-1000	19 X 10^{-7}	0.40
"	"	"	- 500	"	0.24
"	"	"	- 100	"	0.11
MAX.POSITIVE	$\overline{\partial J/\partial z}$	-50	-1000	263 X 10^{-7}	1.32
" "	"	"	- 500	"	1.16
" "	"	"	- 100		1.02
AVE. VALUE	$\overline{\partial J/\partial z}$	-50	- 200	30 X 10^{-7}	0.2

Values used in calculations:

1) $(\rho c)_f$ = 1507 J m^{-3} K^{-1}

2) k_{dry} = 0.025 W m^{-1} K^{-1}

3) L_s = 2.834 X 10^6 J kg^{-1}

4) $\partial T/\partial t$ assumed to be 0

5) ϕ = 0.8

z, the difference between the temperature of the air in the hole (as a function of radius) and the ice is given by (Appendix VI):

$$\text{(5-11)} \qquad \Delta T(r) \approx \left\{1-\frac{r^2}{R^2}\right\}\left\{\frac{v_z R^2 \frac{\partial T}{\partial z}}{4\alpha_{air}}\right\}$$

where r is the radial coordinate, and α_{air} is the thermal diffusivity of air. Applying this model to the Fairbanks snow cover, pore spaces have radii of several millimeters, vertical temperature gradients average -50 K m^{-1}, and the estimated flow velocity is less than or equal to 1.3 mm s^{-1}. The maximum difference in temperature between the center and the walls of the pore would be 0.02°C, and it is unlikely, given this minute temperature difference, that there would be significant under- or over-saturation. Therefore, we conclude that it is satisfactory to assume that the air in the pores is saturated, even when the air is convecting.

For mathematical simplicity, the data for the saturation water vapor density (List, 1951) were fit with an exponential function (Fig. 1-4):

$$\text{(5-12)} \qquad \rho_v(T) = A\, e^{B\,T}$$

where A = 0.00579 and B = 0.09658 for T in °C and ρ_v in kg m^{-3}. Using Equation (5-10) and the relationship $\frac{\partial \rho_v}{\partial z} = \frac{\partial \rho_v}{\partial T}\frac{\partial T}{\partial z}$, Equation (5-6) becomes:

$$(5\text{-}13)\quad \frac{\partial \rho}{\partial t}\Big]_v = \phi\, D_o \Big\{ F \Big[\frac{\partial \rho_v}{\partial T} \frac{\partial^2 T}{\partial z^2} + \Big[\frac{\partial T}{\partial z}\Big]^2 \frac{\partial^2 \rho_v}{\partial T^2} \Big] + \frac{\partial \rho_v}{\partial T} \frac{\partial T}{\partial z} \frac{\partial F}{\partial z} \Big\}$$

$$-\phi \frac{\partial \rho_v}{\partial T} \frac{\partial T}{\partial t} - \phi \Big[\rho_v \frac{\partial v_z}{\partial z} + v_z \frac{\partial \rho_v}{\partial T} \frac{\partial T}{\partial z} \Big]$$

where $\frac{\partial \rho_v}{\partial T} = A\ B\ e^{B\ T}$ and $\frac{\partial^2 \rho_v}{\partial T^2} = A\ B^2\ e^{B\ T}$.

Several terms which appear in Equation (5-13) are not known. These are F, $\frac{\partial F}{\partial z}$, and $\frac{\partial v_z}{\partial z}$. The enhancement factor, F, accounts for the tortuosity of the pore spaces in the snow and the enhanced temperature gradients which result from snow grain geometry. The tortuosity and enhanced gradients have never been measured, yet enhancement factors have been introduced in several mass transport models for snow in order to bring calculated results into agreement with measurements (see Section 1.2.1). Most authors set F equal to approximately 5; however, F should not be a constant since it is a function of the snow grain geometry. It should change with snow density, texture, and perhaps also temperature. However, over small distances, such as the thickness of a snow layer, we will assume it is constant and that $\frac{\partial F}{\partial z}$ is zero. A similar argument can be made for $\frac{\partial v_z}{\partial z}$. With these simplifications, Equation (5-13) can be combined with Equation (5-5) to give:

$$(5\text{-}14)\quad k_{dry}\frac{\partial^2 T}{\partial z^2} + L_s D_o \phi\, F\{AB\,e^{BT}\frac{\partial^2 T}{\partial z^2} + (\frac{\partial T}{\partial z})^2\, AB^2 e^{BT}\} -$$

$$\phi\, v_z \frac{\partial T}{\partial z}\{L_s AB\,e^{BT} + (\rho c)_f\} = \frac{\partial T}{\partial t}\{[(1-\phi)(\rho c)_s + \phi(\rho c)_f] + L_s\phi\, AB\,e^{BT}\}$$

which can be solved for v_z.

To evaluate Equation (5-14), sequential pairs of vertical temperature profiles were fit with quadratic equations using least squares regression. The fitting procedure was used in order to smooth the profiles. From the pairs, T, $\frac{\partial T}{\partial z}$, and $\frac{\partial^2 T}{\partial z^2}$ were determined at 0.05 m, 0.10 m, 0.15 m, 0.20 m, 0.25 m height in the snow. From the change between the first and the second profile, $\frac{\partial T}{\partial t}$ was determined at each height. The air flow velocity was then determined as a function of F.

The average velocity calculated from Equation (5-14) was 0.4 mm s^{-1}, and the maximum was 2 mm s^{-1}. Velocities calculated for coherent and incoherent events did not differ significantly. The results also indicated that there was no simple relationship between the velocity and the choice of the enhancement factor. Increasing F increased the velocity at the base of the snow cover, but decreased it at the top. For example, increasing F from 0.1 to 20 increased the velocity 1000% in the bottom 0.05 m of the snow, but decreased it 10% in the top 0.05 m. The model was sensitive to small uncertainties in $\frac{\partial T}{\partial z}$ and $\frac{\partial^2 T}{\partial z^2}$, suggesting that velocities less than 0.3 mm s^{-1} may not have been significant.

The results of both methods of estimation are consistent: they indicate that the convective flow velocity in the Fairbanks snow cover averaged 0.2 mm s^{-1}, with peak velocities approximately 2 mm s^{-1}.

5.4 Is Convection Important in Moving Heat and Mass?

Heat Transport: During a normal winter, about one third of the heat transported from the base to the top of the Fairbanks snow cover is moved by convection of sensible heat. One third is transported by conduction through the ice grains and air spaces. The remaining third is transported as latent heat with the mass flux. However, during periods when the convective velocity is high (about 2 mm s^{-1}), convection can dominate the heat transport. As a result, during periods of convection, the time-response of the snow to changes in air temperature can be more than an order of magnitude faster than would occur by conduction alone (Table 5-2).

The relative contribution of the latent heat transport can be estimated as follows: heat flow measured by heat flow meters at the base of the snow averaged 10 W m^{-2} between 1985 and 1987 (Fig. 3-16). It is assumed that this is approximately equal to the total heat flow through the snow. The mass flux gradient from the basal half to the top of half of the snow averaged 3 X 10^{-6} kg $m^{-2}s^{-1}m^{-1}$ (see Section 4.2.1), and the average snow depth was 0.6 m (Fig. 3-17). If it is assumed that mass flux out the top of the snow was negligible (see Section 4.2.2) with most of the flux being deposited in the upper part of the snow, then the average mass flux from the base to the top was

$9 \times 10^{-7} kg\ m^{-2} s^{-1}$, though the actual flux gradient is a strong function of depth When multiplied by the latent heat of sublimation, the mass flux resulted in a latent heat flow at mid-height in the snow cover of approximately 2.5 W m^{-2}, which is 26% of the total heat flow. The theoretical estimates of Yosida et al. (1955), Woodside (1958), and de Quervain (1972) for the contribution of vapor diffusion to the total thermal conductivity ranged from 40% to 48% (see Section 1.2.2).

The estimate of the importance of the latent heat transport can be corroborated by comparing the magnitudes of the three heat transport terms in Equation (5-5). The ratio of the latent heat term to the sum of the magnitudes of the conductive, convective and latent heat terms can be written:

$$\frac{\frac{\partial \overline{J}}{\partial z} L_s}{\phi\ (\rho c)_f\ \overline{v}_z \frac{\partial \overline{T}}{\partial z} + k_{dry} \frac{\partial^2 \overline{T}}{\partial z^2} + \frac{\partial \overline{J}}{\partial z} L_s} \tag{5-15}$$

This ratio was evaluated from the data in Table 5-4. The ratio ranged from 0.04 to 0.5, but for reasonable values of the curvature ($\leq -200°C\ m^{-2}$), the range was 0.3 to 0.5, which is consistent with the previous estimate that a third of the total heat was transported as latent heat.

The relative importance of conduction to convection of dry air (sensible heat) can be estimated using the Peclet number (Pe_h). It is defined as the ratio of the convective to the conductive heat transport under conditions where there is no latent heat transport, and can be written:

$$Pe_h = \{\frac{h\ v_z\ \phi}{\alpha_{dry}}\ \frac{(\rho c)_f}{(\rho c)_m}\}. \tag{5-16}$$

Peclet numbers less than 1 indicate that conduction dominates the heat transport, while numbers greater than 1 indicate convection dominates.

Peclet numbers, calculated using the velocity estimates from Sections **5.3.2** and **5.3.3**, indicate that convection and conduction were usually of equal importance in transporting heat, except during the few periods when the convective velocity was high (~ 2 mm s^{-1}) (Table 5-5). For example, at the average velocity of 0.2 mm s^{-1}, Pe_h ranged 0.8 to 1.6, depending on snow depth, but for the maximum velocity of 2 mm s^{-1}, it ranged from 8 to 17.

Mass Transport: The contribution of convection to the water vapor transport can be estimated by examining the ratio of the convective to the diffusive transport terms in Equation (5-13). This ratio is:

$$\frac{v_z \frac{\partial T}{\partial z}}{D_o F \{\frac{\partial^2 T}{\partial z^2} + B \frac{\partial T}{\partial z}^2\}} \tag{5-17}$$

For the average velocity (0.2 mm s^{-1}) and the commonly used value of the enhancement factor (F = 5; see Section **1.2.1**), the ratio is 0.2 (Table 5-6). Even through a wide range of values of the enhancement

Table 5-5: Peclet numbers for heat transport. v_z is velocity, h is snow depth. Peclet numbers for the average velocity are underlined.

v_z (mm/s)	h (m)	Pe_h
0.01	0.15	.05
0.1	0.15	.4
0.2	0.15	0.8
1.0	0.15	4.2
2.0	0.15	8.4
0.01	0.30	.1
0.1	0.30	0.8
0.2	0.30	1.6
1.0	0.30	8.5
2.0	0.30	17.0

Table 5-6: The ratio of convective to diffusive mass transport. The value of F is poorly known; a value of F = 5 has been suggested by several authors (see Section 1.2.1). The estimated average velocity for the Fairbanks snow cover is 0.2 mm s^{-1}, and the maximum is 2 mm s^{-1} (see Sections 5.3.2 and 5.3.3).

v_z (mm/s)	F	$\frac{\partial T}{\partial z}$ (K m^{-1})	$\frac{\partial^2 T}{\partial z^2}$ (K m^{-2})	$\frac{\text{CONVECTION}}{\text{DIFFUSION}}$
0.2	1	-50	-500	1.8
0.2	2	-50	-500	1.0
0.2	5	-50	-500	0.4
0.2	10	-50	-500	0.2
2.0	1	-50	-500	20.0
2.0	2	-50	-500	10.0
2.0	5	-50	-500	4.0
2.0	10	-50	-500	1.8

factor ($2 \leq F \leq 10$), the ratio is no greater than 1. This suggests that diffusion is the primary agent of water vapor transport in the Fairbanks snow cover most of the time. However, for the maximum estimated convective velocity (2.0 mm s^{-1}), the ratio is one to two orders of magnitude greater, ranging from 2 to 20, depending on the value of F. When F equals 5, the ratio equals 4. This indicates that considerable water vapor can be transported by convection during periods of peak flow.

Recalling the results of Section 4.2.2, it was suggested that a significant percent of the layer-to-layer mass transport occurred in brief, intense intervals of enhanced flux. One might speculate that these occurred during periods of extremely low air temperatures, when temperature gradients across the snow were maximum. Under these conditions, the diffusive vapor transport is certainly enhanced, since it is a function of the temperature gradient and the curvature of the vertical temperature profile. But these low temperature periods were also times when strong incoherent events not associated with wind occurred. Potentially, the convective flow velocity was high during these periods. In that case, the ratio in (5-17) may have been significantly greater than 1, and convection could have increased the already enhanced diffusive vapor flux by a factor of 10 or more. The contribution of these intense periods of convective vapor transport to the total vapor transported during the winter would depend on how often they occurred. If they occurred each time there was an incoherent event (about 15% of the winter), they would make a significant contribution. Unfortunately, the actual prevalence of higher velocity convective flow could not be determined from the data collected in the present study.

The virtual absence of temperature profiles with positive curvature, out of the thousands which were measured, suggests that diffusion is the primary agent of water vapor transport, and implies that periods when convention moves significant amounts of water vapor are infrequent. The negative curvature of the vertical temperature profiles was discussed in Section **3.2.3**. It could be either 1) the result of upward diffusion of vapor and the liberation of latent heat, and/or 2) the result of upward advection of warm air. Advection of warm air, however, can cause either negative or positive curvature, depending on whether the flow direction is up or down. Based on the hypothetical circulation pattern (Fig. 5-1), the probability of making a temperature profile measurement in an updraft or downdraft is nearly equal. On the other hand, the upward diffusion of water vapor will always yield negative curvature in the temperature profile. This suggests that the negative curvature was generally the result of the upward diffusion of vapor.

Though convection may be generally unimportant in moving water vapor, it may a play in role in the development of the snow textures described in Section **3.1.3**. There is a striking coincidence between the location of Metamorphic Layer 5, in which the predominant c-axis orientations are horizontal, and the zone at the base of the snow where convective flow lines are thought to be horizontal (Figs. 3-10, 5-1). Elsewhere in the snow, flow lines are predominantly vertical, as are c-axes. Perhaps the air flow direction in the snow influences the growth direction of the crystals.

6 SUMMARY OF RESULTS, CONCLUSIONS, AND RECOMMENDATIONS FOR FUTURE WORK

The results of the present study demonstrate that convection occurs in a natural snow cover, but the same results raise many questions. In particular, what is the prevalence of convection in other types of snow covers, what is its relationship to snow metamorphism, and what conditions are necessary for the onset of convection? A summary of results and conclusions (below) is followed by recommendations for studies that address these questions.

6.1 Summary of Results and Conclusions

The Fairbanks snow progressively metamorphoses, until by the end of the winter it consists almost entirely of depth hoar. A five-layer sequence develops with a basal layer composed of extremely large crystals (≥ 10 mm) with horizontal c-axes; it is overlain by layers composed of crystals with vertical c-axes. The metamorphism results in a rapid increase in air permeability to values two to three times higher than previously measured, and a decrease in the thermal conductivity to the lowest previously measured value. Also, the number of grains per unit volume decreases an order of magnitude as a few select grains grow, and the rest of the grains shrink by sublimation.

The metamorphism is the result of strong temperature gradients caused by relatively high snow-soil interface temperatures and low air temperatures. These temperature gradients drive vapor diffusion and convection which cause a net transfer of mass from the base to the top of the snow. Over the past twenty years, the average layer-to-layer mass flux gradient was 3×10^{-6} kg $m^{-2}s^{-1}m^{-1}$, but there were occasional periods when it was a factor of ten greater. We speculate that these periods of enhanced flux coincide with periods of strong, higher velocity convection.

Thermal convection is common in the Fairbanks snow cover. It occurred sporadically in 1985, but nearly continuously in 1986 and 1987. The evidence for the convection is:

1) simultaneous warming and cooling at different locations in a horizontal plane in the snow,
2) observations of up to a 16 K range in temperature over a horizontal distance of about a meter,
3) response of the snow to rapid changes in air temperature 10 to 50 times faster than can be explained by diffusion, and
4) an average Nussult number (the ratio of the total effective bulk conductivity of the snow measured by heat flow meters to the average effective thermal conductivity) equal to, or greater than 2.

The convection is time-dependent. A quasi-stable mode associated with coherent temperature signals is present virtually throughout the winter. Occasionally (≤ 16% of the winter), this convective flow is perturbed by external conditions such as high wind (wind speed ≥ 5 m

s^{-1}) or rapid changes in temperature. At those times, the convective flow velocity probably increases, creating thermal disequilibrium in the snow, and resulting in incoherent temperature signals.

A plume-like convective flow pattern is suggested by relative warm and cold zones which develop in the snow and remain relatively fixed in space through the winter. The zones seem to be linked to spatial variations in the temperature of the snow-soil interface that result from uneven soil moisture or lateral variations in the thermal conductivity of the soil. By analogy with laboratory experiments of porous media, it is likely that warm, diffuse plume-like updrafts develop in the snow above the regions of the interface which are at higher temperature. The plume-like circulation pattern is characterized by vertical air flow except in a narrow layer at the base of the snow where flow is horizontal.

The onset of convection in the snow cannot be predicted using a standard Rayleigh number formulated for a homogeneous porous medium with isothermal upper and lower boundaries. The critical Rayleigh number formulated in this manner was never exceeded during all three winters of the study, despite the fact that the system was almost continuously convecting. The proper formulation of the Rayleigh number for a system with ever-changing boundary conditions and a non-uniform basal temperature is not known. Spatial variations in the temperature of the snow-soil interface, the continuously fluctuating air temperature, and on-going snow metamorphism probably all contribute to a dramatic lowering of the critical threshold for the initiation of convection.

Estimates based on one-dimensional heat and mass transport equations suggest that the average convective velocity in the Fairbanks snow cover is on the order of 0.2 mm s^{-1}, with a maximum value of 2 mm s^{-1}. Using this range of velocities, it is estimated that convection accounts for a third of the total heat transport. Latent heat transported by the vapor flux accounts for another third. Most of the time the vapor flux is primarily the result of diffusion, but during the brief periods when the convective flow velocity approaches its maximum value, it can increase the mass transport by a factor of 10 or more. During these same periods, convection of dry air moves up to 17 times as much heat as is conducted through the ice matrix and air spaces.

Convection may be important in the development of the five-layer metamorphic sequence. The basal layer, which consists of crystals with horizontal c-axes, coincides with the zone where the air flow is thought to be horizontal. The overlying layers, which contain crystals with vertical c-axes, coincide with vertical flow, suggesting that the flow direction affects the crystal growth.

The spatial variations in temperature at the snow-soil interface probably lowered the critical Rayleigh number for the onset of convection at the experimental site, which was carefully leveled and bare of vegetation. Natural, vegetated snow-soil interfaces undoubtedly develop greater temperature contrasts than observed in this study. These greater contrasts would promote convection.

Forced convection due to the action of the wind is probably more likely in other snow covers. In the present study 50% of all incoherent events coincided with windy periods, yet the maximum wind speed recorded was only 10 m s^{-1}. Several of the events coincided with wind speeds as low as 5 m s^{-1}. Most snow-covered regions are windier than Fairbanks, making forced convection a strong possibility in these locations.

6.2 Recommendations for Future Work

6.2.1 Determine the Prevalence of Convection

1) Determine the extent of convection in other types of snow: One vertical and one horizontal thermistor string could be installed and monitored in several different types of snow covers. These must be installed over carefully prepared substrates, but they would be sufficient to indicate the presence of convection. The measurements could be made in several extreme types of snow covers, for example in the deep, dense snow found in Maritime climates, in the thin, wind-blown snow found on the arctic coastal plain, and in the snow found on sea ice.

2) Determine the extent of convection on hillsides: In theory (Bories and Combarnous, 1973; Powers et al., 1985) air in the snow on hillsides always convects. An installation of thermistors in several different types of snow covers on hillsides should be able to verify this and help define the extent and characteristics of the convection.

3) Determine the convective velocity: Convective velocities could be measured using tracer gases. Sulphur-hexafluoride or another trace gas could be introduced at the bottom of the snow during convection

events by installing breakable glass vials containing the gas at the soil surface prior to the first snowfall. During the events, the vials could be broken remotely and the time it took the gas to travel from the vials to receptor tubes could be measured and used to calculate the air flow velocity. An experiment of this nature was designed and installed during the present study but never run successfully.

4) Measure temperature variations at the snow-soil interface: Spatial variation in the temperature of the snow-soil interface should be measured for different substrates and moisture conditions. Because these variations affect the onset and presence of convection, it would be useful to know their range, and how they change with vegetation, micro-topography and moisture content.

6.2.2 Studies of the Relationship between Convection and Snow Metamorphism

5) Create an artificial warm zone at the base of the snow: An artificial warm zone could be created at the base of a natural snow cover using a controllable line heater placed on the ground before the first snowfall. The heater could be adjusted during the winter to maintain it at a higher temperature than the surrounding soil surface, and at the end of winter, the snow texture above the heater and elsewhere could be compared.

6) Grow depth hoar with and without air flow: In a laboratory, natural new snow samples could be subjected to temperature gradients

similar to those found in the Fairbanks snow. Inverted gradients (warm on top, cold on the bottom) could be applied to some samples to prevent convection, while normal gradients could be applied to other samples, allowing convection. Controlled ventilation of some samples could be used to try to determine the effect of air flow on the c-axes orientation in growing crystals.

6.2.3 Studies of the Onset of Convection

7) Induce convection by creating a warm zone at the base of the snow: One of the samples used in (6) could be instrumented with thermistors and placed on a plate which had a controllable heat pad in one section. The onset of convection in the snow could be investigated at different temperature gradients and at different stages of depth hoar metamorphism.

8) Develop onset criteria for porous media over a warm spot: Using existing computer models for convection in porous media, it should be possible to investigate the onset criteria for porous media with a non-uniform basal boundary temperature. It would also be useful to include in the models the effects of latent heat transfer, and rapidly varying temperatures at the upper surface.

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APPENDIX I: AIR PERMEABILITY MEASUREMENTS

The air permeability measurements were made by E. Chacho. Limited measurements were made in 1985 and 1986. The bulk of the measurements were made in 1987 using a new air permeameter designed, built, and described by Chacho and Johnson (1987). Figure I-1 illustrates the increase in permeability during the winter as a result of snow metamorphism. The results for 1987 are listed in Table I-1. Vertical and horizontal core samples from the snow were measured and are listed separately.

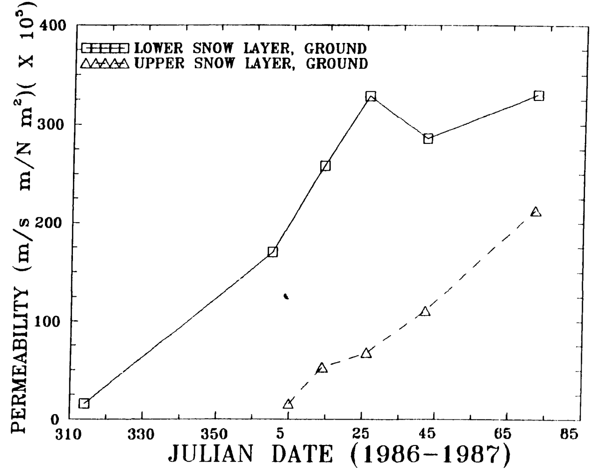

FIGURE I-1: The air permeability of two layers of snow on the ground, 1987. The lower layer was deposited on Julian date 311; the upper layer was deposited between Julian date 350 and 370. Measurements were made on vertical core samples.

TABLE I-1: Air permeability measurements for the Fairbanks snow, 1987. Vertical and horizontal core of snow were measured.

ground

DATE 1987	SAMPLE	DENSITY (kg/m^3)	PERMEABILITY (m/s)*10^-5 (m/N*m^-2)
UPPER SNOW LAYER			
VERTICAL			
5-Jan	4	100	16
14-Jan	5	160	53
26-Jan	14	170	68
11-Feb	24	180	111
13-Mar	38	180	213
HORIZONTAL			
14-Jan	10	160	69
26-Jan	19	170	90
11-Feb	30	160	54
13-Mar	43	170	109
LOWER SNOW LAYER			
VERTICAL			
31-Dec	3	165	170
14-Jan	7	175	258
26-Jan	16	175	329
11-Feb	26	180	286
11-Feb	27	180	314
13-Mar	41	180	287
13-Mar	39	180	330
HORIZONTAL			
14-Jan	9	150	198
26-Jan	20	160	313
11-Feb	31	175	310
13-Mar	44	170	335
TOTAL SNOW PACK			
14-Jan	8		105
26-Jan	18		172
11-Feb	28		172
11-Feb	29		186
13-Mar	42		213

tables

DATE 1987	SAMPLE	DENSITY (kg/m^3)	PERMEABILITY (m/s)*10^-5 (m/N*m^-2)
UPPER SNOW LAYER			
VERTICAL			
14-Jan	11	110	46
26-Jan	21	125	44
11-Feb	32	150	45
13-Mar	45	170	62
HORIZONTAL			
11-Feb	35	150	79
13-Mar	48	145	46
LOWER SNOW LAYER			
VERTICAL			
31-Dec	1	200	38
31-Dec	2	200	33
14-Jan	13	200	46
26-Jan	22	220	44
11-Feb	33	210	50
13-Mar	46	250	47
HORIZONTAL			
13-Mar	49	230	38
TOTAL SNOW PACK			
14-Jan	12		57
26-Jan	23		42
11-Feb	34		49
13-Mar	47		39

APPENDIX II: INSTRUMENTATION ERRORS

The data logger system could resolve ± 1 ohm when set at a full scale reading of 30 K ohms. Age and operating temperature of the logger reduced the accuracy by ± 0.016% of the resistance reading, which normally ranged between 5000 and 45000 ohms. Maximum instrumentation error was ± 0.02°C. Excitation current supplied by the logger was 100 μA. 250 thermistors were scanned in less than 180 seconds, including switching time; current was supplied to each thermistor for less 0.5 s. An estimate of the possible self heating of thermistors due to this excitation current can be determined as follows: Maximum self heating and change in temperature will occur if the thermistors are insulated (adiabatic system). In nature, conduction of heat to the environment occurs, reducing this change in temperature. The power input to the thermistor is:

$$\text{(A-1)} \qquad P = i^2 \, \Omega$$

where P is the power, i is the current, and Ω is the thermistor resistance. Let c_t be the specific heat capacity of the thermistor, and m_t be the mass of the thermistor. For the adiabatic case:

$$\text{(A-2)} \qquad \frac{dT}{dt} = [P/(m_t \, c_t)]$$

or integrating from time 1 to time 2:

$$\text{(A-3)} \quad \Delta T = [P/(m_t \; c_t)] \; t$$

Using measured values ($t = 0.5$ s; $m_t = 3 \times 10^{-6}$ kg;

$c_t = 800$ J kg^{-1} °C^{-1}; $i = 100$ μA, $\Omega = 5000$ ohms),

$$\text{(A-4)} \quad \Delta T \leq 0.01°C$$

For the thermistors used in this study, a conservative estimate of the maximum error due to Joule heating and instrumentation drift is ± 0.03°C.

The data logger system could resolve 10 μvolts. The only transducers for which DC voltages were recorded were the heat flow meters. For these, instrument error was about ± 10 μV, which gives rise to errors in heat flow of 0.3 W m^{-2} which are insignificant compared the order of magnitude of the results (5 to 30 W m^{-2}).

APPENDIX III: THERMAL CONDUCTIVITY MEASUREMENTS

The thermal conductivity was measured using a needle probe method which is based on the rate of warming or cooling of a sample during the transient heating of a line source embedded in the sample. It has been described by Blackwell (1954), Lachenbruch (1957), Jaeger (1958), von Herzen and Maxwell (1959), Pratt (1969), McGaw (1984), and Jones (1988), and has been used to measure the thermal conductivity of snow by Jaafar and Picot (1970), Lange (1985), and Sturm and Johnson (1987).

A stainless steel needle probe built by Custom Scientific Instruments Inc. (Model CS-48/E), 0.2 m long and containing a helical heating wire and a Chromel-Constantan thermocouple, was inserted along the axis of a 0.10 m diameter cylindrical snow sample which had equilibrated to $\pm$ 0.01°C at the desired temperature using a copper jacket and circulating bath. Current (using 10 to 16 volts) was fed to the heater for a 5 to 10 minute heating phase, then turned off. The sample was then allowed to cool for 15 to 20 minutes. During both heating and cooling phases, the temperature of the needle was monitored. From the temperature data, a separate value of the thermal conductivity could be calculated for both the heating and cooling phase (McGaw, 1984). If they agreed to $\pm$ 10%, the test was considered accurate and the results averaged. If not, the test was repeated after the sample had re-equilibrated.

The tests were run using an Hewlett Packard 41-CV calculator and 3421A data logger in 1985 and 1986. In 1987, the data logger and power supply was controlled by a Hewlett Packard Integral Computer.

The results of all measurements made between 1984 to 1987 are listed in Table III-1. The thermal conductivity measured during the heating phase (k_h) and the thermal conductivity measured during the cooling phase (k_c) were averaged. This average is listed in Table III-1 as k_{ave}.

TABLE III-1: Summary of thermal conductivity measurements.

DATE	SNOW TYPE	DENSITY [kg m^{-3}]	TEMP. [°C]	HT. [m]	k_h [watts $m^{-1}K^{-1}$]	k_c	k_{ave}	GOOD TEST?
1985–1986								
Nov. 16	New snow	98	-13.6	--	0.049	0.098	0.049	No
Nov. 16	"	98	-5.5	--	0.039	--	--	(?)
Nov. 16	New snow	98	-8.2	--	0.047	--	--	(?)
Nov. 24	Depth hoar	179	-6.8	0.03	0.031	--	0.031	Yes
Nov. 24	"	"	-10.1	"	0.017	--	(?)	No
Nov. 24	"	"	-7.0	"	0.030	0.031	0.031	Yes
Nov. 24	"	162	-4.8	"	0.034	--	0.034	(?)
Nov. 24	Depth hoar	179	-7.5	"	0.049	0.115	0.049	No
Jan. 30	Depth hoar	183	-24.9	0.15	0.045	--	0.045	Yes
Jan. 30	"	"	-27.0	"	0.048	0.068	0.058	(?)
Jan. 30	Depth hoar	183	-21.8	0.15	0.132	--	(?)	No
Feb. 1	Depth hoar	180	-26.1	0.15	0.067	0.101	0.084	(?)
Mar. 6	Depth hoar	195	-14.6	0.15	0.050	0.055	0.053	Yes
Mar. 6	"	"	-13.4	"	0.057	--	0.057	(?)
Mar. 6	Depth hoar	195	-12.6	0.15	0.063	--	0.063	Yes
1986–1987								
Dec. 20	Depth hoar	160	-19.2	0.10	0.053	--	0.053	Yes
Dec. 20	"	"	-19.3	"	0.049	--	0.049	Yes
Dec. 20	"	"	-19.1	"	0.054	--	0.054	Yes
Dec. 20	"	"	-6.4	"	0.053	--	0.053	Yes
Dec. 20	"	"	-6.4	"	0.054	0.055	0.054	Yes
Dec. 20	"	"	-19.3	"	0.059	0.058	0.058	Yes
Dec. 20	Depth hoar	160	-19.8	0.10	0.064	0.067	0.066	Yes
Dec. 31	Depth hoar	175	-10.3	0.03	0.065	0.060	0.063	Yes
Dec. 31	"	"	-19.8	"	0.043	--	0.043	(?)
Dec. 31	"	"	-10.3	"	0.049	0.056	0.052	Yes
Dec. 31	"	"	-18.6	"	0.049	0.065	0.057	(?)
Dec. 31	"	"	-19.8	"	0.043	0.053	0.048	Yes
Dec. 31	"	175	-10.2	0.03	0.052	0.057	0.054	Yes

TABLE III-1 cont.

DATE	SNOW TYPE	DENSITY [kg m^{-3}]	TEMP. [°C]	HT. [m]	k_h	k_c	k_{ave} [watts m^{-1}K^{-1}]	GOOD TEST?
1986-1987								
Jan. 6	New snow	70	-19.6	0.30	0.058	0.054	0.056	Yes
Jan. 6	New snow	70	-19.2	0.30	0.060	0.054	0.057	Yes
Jan. 14	Depth hoar	155	-4.9	0.07	0.049	0.048	0.049	Yes
Jan. 14	"	"	-18.8	"	0.049	0.050	0.049	Yes
Jan. 14	"	"	-18.9	"	0.049	0.043	0.046	Yes
Jan. 14	Depth hoar	155	-18.9	0.07	0.050	0.053	0.047	(?)
Jan. 14	Depth hoar	195	-19.2	0.17	0.081	0.078	0.079	Yes
Jan. 14	"	"	-19.0	"	0.092	0.078	0.085	(?)
Jan. 14	"	"	-19.0	"	0.096	0.085	0.091	(?)
Jan. 14	"	"	-19.3	"	0.086	0.079	0.082	Yes
Jan. 14	"	"	-5.0	"	0.105	0.099	0.102	Yes
Jan. 14	Depth hoar	"	-5.0	"	0.098	0.098	0.098	Yes
Jan. 26	Depth hoar	168	-18.4	0.07	0.053	0.051	0.052	Yes
Jan. 26	"	"	-18.4	"	0.057	0.046	0.052	(?)
Jan. 26	"	"	-18.6	"	0.052	0.049	0.051	Yes
Jan. 26	"	"	-12.3	"	0.080	0.084	0.082	Yes
Jan. 26	Depth hoar	168	-12.4	0.07	0.077	0.080	0.078	(?)
Jan. 26	Depth hoar	201	-19.7	0.17	-	-	-	-
Jan. 26	"	"	-19.6	"	0.078	0.068	0.073	(?)
Jan. 26	"	"	-19.5	"	0.093	0.071	0.082	No
Jan. 26	"	"	-19.5	"	0.084	0.082	0.083	Yes
Jan. 26	"	"	-11.1	"	0.057	0.054	0.055	(?)
Jan. 26	"	"	-11.1	"	0.069	0.056	0.063	No
Jan. 26	Depth hoar	201	-4.8	0.17	0.079	0.090	0.085	No
Jan. 26	Old Stellars	169	-19.2	0.29	0.059	0.055	0.057	Yes
Jan. 26	"	"	-19.3	"	0.061	0.055	0.058	(?)
Jan. 26	"	"	-18.9	"	0.056	0.056	0.056	Yes
Jan. 26	Old	"	-12.9	"	0.063	0.057	0.060	Yes
Jan. 26	Stellars	"	-12.8	"	0.062	0.055	0.058	(?)
Feb. 10	Depth hoar	170	-19.2	0.07	0.045	0.042	0.044	Yes
Feb. 10	Depth hoar	170	-18.5	0.07	0.043	0.046	0.044	Yes

TABLE III-1 cont.

DATE	SNOW TYPE	DENSITY [kg m^{-3}]	TEMP. [°C]	HT. [m]	k_h [watts m^{-1}K^{-1}]	k_c	k_{ave}	GOOD TEST?
1986-1987								
Feb. 10	Depth hoar	200	-20.0	0.17	0.071	0.059	0.065	(?)
Feb. 10	Depth hoar	168	-19.9	0.27	0.063	0.065	0.064	Yes
Mar. 13	Depth hoar	154	-19.6	0.04	0.028	0.023	0.021	No
Mar. 13	"	"	-19.8	"	0.076	0.069	0.072	Yes
Mar. 13	"	"	-19.7	"	0.076	0.071	0.073	Yes
Mar. 13	Depth hoar	154	-19.9	0.04	0.079	0.074	0.077	Yes
Mar. 13	Depth hoar	190	-20.1	0.17	0.093	0.086	0.090	Yes
Mar. 13	"	"	-20.1	"	0.086	0.083	0.084	Yes
Mar. 13	"	"	-20.1	0.17	0.093	0.085	0.089	Yes
Mar. 13	Depth hoar	190	-20.2	0.25	0.096	0.086	0.091	Yes
Mar. 13	Depth hoar	160	-20.3	0.31	0.086	0.092	0.089	Yes
Mar. 26	Depth hoar	160	-18.2	0.05	0.056	0.051	0.053	Yes
Mar. 26	"	"	-18.1	"	0.055	0.050	0.053	Yes
Mar. 26	"	"	-18.5	"	0.057	0.049	0.053	No
Mar. 26	Depth hoar	160	-18.8	0.05	-	-	-	-
Mar. 26	Depth hoar	220	-18.9	0.16	0.042	0.038	0.040	Yes
Mar. 26	"	"	-19.1	"	0.045	0.040	0.042	Yes
Mar. 26	Depth hoar	220	-19.2	0.16	0.045	0.044	0.044	Yes
Mar. 26	Depth hoar	190	-18.5	0.26	0.070	0.056	0.063	(?)
Mar. 26	"	"	-18.4	"	0.064	0.061	0.062	Yes
Mar. 26	"	"	-11.5	"	0.075	0.066	0.070	(?)
Mar. 26	"	"	-6.7	"	0.083	0.074	0.079	Yes
Mar. 26	"	"	-6.7	"	0.084	0.081	0.083	Yes
Mar. 26	Depth hoar	190	-3.6	0.26	0.089	0.074	0.082	(?)
FOAM STANDARD								
Jan. 2/87	FOAM	-	ROOM	-	0.009	0.013	0.011	No
Jan.14/87	FOAM	-	ROOM	-	0.010	0.012	0.011	Yes
Feb. 5/87	FOAM	-	ROOM	-	0.012	0.014	0.013	Yes
Mar.16/87	FOAM	-	ROOM	-	0.012	0.011	0.011	Yes
Mar.16/87	FOAM	-	ROOM	-	0.012	0.015	0.014	(?)
May 15/87	FOAM	-	ROOM	-	0.013	0.015	0.014	Yes
Jul.21/88	FOAM	-	ROOM	-	0.014	0.017	0.016	(?)

APPENDIX IV: CALCULATION OF THE NUMBER OF GRAINS IN A SNOW SAMPLE

Let a snow sample be sieved through L sieves, the bottom-most sieve actually being a pan which catches all the grains that passed through the immediately larger sieve. The total mass in the j^{th} sieve, M_j, is equal to the sum of the masses of all the individual grains in the sieve:

$$(A\text{-}1) \qquad M_j = \sum_{i=1}^{N_j} m_i$$

where there are N_j grains in the j^{th} sieve, and the mass of the i^{th} grain contained in the sieve is m_i. If the average mass of a grain in the j^{th} sieve is $\bar{m}_j$, then:

$$(A\text{-}2) \qquad M_j = N_j \, \bar{m}_j$$

By weighing individual grains in each sieve, a curve of average grain mass vs. sieve size was established (Fig. 3-8). From this curve, values of $\bar{m}_j$ can be determined for any sieve size, which makes it possible to calculate the number of grains in the j^{th} sieve using Equation (A-2).

The total number of grains in the sample, N_T, can be calculated by adding up the number of grains in each sieve:

$$N_T = \sum_{j=1}^{L} N_j. \tag{A-3}$$

In this study L equaled 9, and all samples were of a constant mass, M, rather than a constant volume. Therefore, N_T is the total number of grains per sample mass. The total number of grains per unit volume is:

$$N = N_T \frac{\rho}{M} \tag{A-4}$$

where ρ is the density of the snow sample.

The uncertainties in N were large. The uncertainty in the value of $\bar{m}_j$ was probably ± 20%, producing uncertainties in N of ± 20%. Much of total uncertainty in N was the result of enormous number of grains in the pan at the bottom of the sieve stack, which caught all the grain fragments. The pan could contain as many as 3 X 10^8 grains per cubic meter, with a corresponding uncertainty of ± 6 X 10^7 grains per cubic meter. These uncertainties dominated the total uncertainty when there was a significant fraction of the sample in the pan. Generally, however, there was a relatively small number of grains in the pan and the corresponding uncertainty was smaller.

Some of the abrupt, large changes in N for the samples from the tables (Fig. 3-9) may have had a physical cause. Relict layers of surface hoar buried by other snow were encountered in sampling. Where possible, samples unbiased by surface hoar were preferred, but it is possible that occasional samples were contaminated with large surface hoar grains, which would have had the effect of markedly decreasing N.

APPENDIX V: THE CONTINUITY EQUATION FOR A COMPACTING LAYER OF SNOW

Conservation of mass for a small volume of snow, ΔV, from which there is mass flux divergence, $\nabla \cdot \vec{J}$, can be written as:

$$\text{(A-1)} \qquad \lim_{\Delta V \to 0} \left\{ \frac{1}{\Delta V} \frac{Dm}{Dt} \right\} = -(\nabla \cdot \vec{J})$$

where t is time and $\frac{Dm}{Dt}$ is the material derivative which is used because the volume is moving. The flux divergence, $\nabla \cdot \vec{J}$, is primarily the result of the movement of water vapor into and out of the volume by diffusion and convection. If we set the mass equal to $(\rho \, \Delta V)$, Equation (A-1) can be written:

$$\text{(A-2)} \qquad \lim_{\Delta V \to 0} \left| \rho \frac{1}{\Delta V} \frac{D(\Delta V)}{Dt} \right| + \frac{D\rho}{Dt} = -(\nabla \cdot \vec{J})$$

The first term in brackets is the density (ρ) times the volumetric strain rate $(\dot{e})$, so Equation (A-2) becomes:

$$\text{(A-3)} \qquad \dot{e}\,\rho + \frac{D\rho}{Dt} = -(\nabla \cdot \vec{J})$$

Assuming that the small volume is in a horizontal snow layer of great extent, there will be no strain in the x and y direction. Then:

$$\text{(A-4)} \qquad \dot{e} \to \dot{e}_{zz} = \lim_{h \to 0} \left\{ \frac{1}{h} \frac{Dh}{Dt} \right\}$$

where h is the thickness of the snow layer. Further, assuming that water vapor flows into or out of this layer only in a vertical direction (the predominant temperature gradients are in that direction), then $\nabla \cdot \vec{J}$ simplifies to $\frac{\partial \bar{J}}{\partial z}$. Equation (A-3) becomes:

$$\text{(A-5)} \qquad \lim_{h \to 0} \left\{ \frac{\rho}{h} \frac{Dh}{Dt} + \frac{D\rho}{Dt} \right\} = - \frac{\partial \bar{J}}{\partial z}$$

If h is finite, then this is:

$$\text{(A-6)} \qquad \frac{\bar{\rho}}{h} \frac{Dh}{Dt} + \frac{D\bar{\rho}}{Dt} \approx - \frac{\partial \bar{J}}{\partial z}$$

where h is the thickness of the layer, and z is the vertical coordinate, positive upward, and the overbar indicates the spatial average over the layer thickness.

The uncertainty in the calculated value of $\frac{\partial J}{\partial z}$ can be determined by considering the uncertainties (e) in the measured values of h, ρ, $\frac{Dh}{Dt}$, and $\frac{D\rho}{Dt}$:

snow layer thickness (h): $e_h = \pm 3$ mm

snow layer density (ρ): $e_\rho = \pm 5$ kg m^{-3}

slope of compaction curve $(\frac{Dh}{Dt})$: $e_{Dh} = \pm 6 \times 10^{-10}$ m s^{-1}

slope of densification curve $(\frac{D\rho}{Dt})$: $e_{D\rho} = \pm 1 \times 10^{-6}$ kg m^{-3}s^{-1}

This produces an uncertainty in the calculated value of $\frac{\partial J}{\partial z}$ equal to (Young, 1962):

$$\text{(A-7)} \quad e = [(\frac{\frac{Dh}{Dt}}{h} e_{\rho})^2 + (\frac{\frac{Dh}{Dt}(\rho)}{h^2} e_h)^2 + (\frac{\rho}{h} e_{Dh})^2 + (e_{D\rho})^2]^{1/2}$$

where h(t) and ρ(t) are the curves fit to compaction and densification data (see Section 4.2). The uncertainty in $\frac{\partial J}{\partial z}$ (using reasonable values of h, ρ, and h(t) for purposes of calculation [h = 0.05 m, ρ = 180 kg m^{-3}, and $\frac{Dh}{Dt} = 50 \times 10^{-10}$ m s^{-1}]) shown as a function of e_h and e_{ρ} in Figure V-1 indicates that the flux gradient is more sensitive to errors in thickness than errors in density.

The flux gradient, $\frac{\partial J}{\partial z}$, can be integrated starting at the snow-soil interface in order to calculate the flux at the top of each snow layer:

(A-8) $$J(z) = \int_0^h \left(\frac{\partial J}{\partial z}\right) dz$$

The flux out of the bottom layer of snow (thickness h_b) is:

(A-9) $$J(h_b) = h_b \frac{\partial J}{\partial z} + J_{soil-snow}$$

where $J_{snow-soil}$ is the flux measured at the snow-soil interface. The integration can be repeated for successively higher layers in the snow cover.

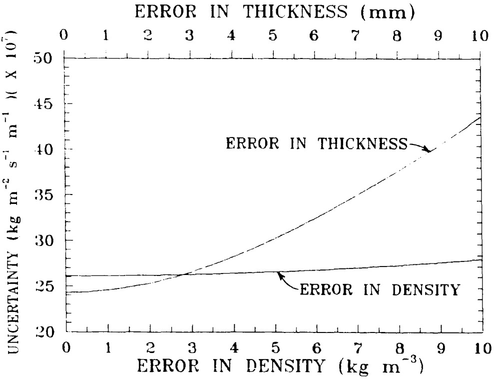

FIGURE V-1: The relationship between the uncertainty (e) in the calculated value of the layer-to-layer mass flux gradient ($\partial J/\partial z$) and errors in measurements of snow density (ρ) and layer thickness (h).

APPENDIX VI: TEMPERATURE DISTRIBUTION IN A CYLINDRICAL HOLE IN AN ICE BLOCK

Consider an ice block of thickness H through which there is a vertical, cylindrical hole of radius R. Air flows up the hole at a velocity, v_z. There is a vertical temperature gradient across the block, $\frac{\partial T}{\partial z}$, where z is the vertical coordinate, positive upward. In steady-state, the heat equation for the moving air in the hole is:

$$\text{(A-1)} \qquad \frac{\partial^2 T}{\partial z^2} + \frac{1}{r}\left\{\frac{\partial}{\partial r} r \frac{\partial T}{\partial r}\right\} + \frac{v_z}{\alpha_{air}} \frac{\partial T}{\partial z} = 0$$

where α_{air} is the thermal diffusivity of the air, and r is the radial coordinate from the center of the hole. Assuming that the temperature gradient is a constant, it can be shown by differentiation that a solution to Equation (A-1) is:

$$\text{(A-2)} \qquad T(z,r) = \frac{\partial T}{\partial z} z + \left\{1 - \frac{r^2}{R^2}\right\}\left\{\frac{v_z R^2 \frac{\partial T}{\partial z}}{4\alpha}\right\}.$$

The difference in the temperature of the air at a height z from that of the walls of the hole is:

$$\text{(A-3)} \qquad \Delta T(r) = \left\{1 - \frac{r^2}{R^2}\right\}\left\{\frac{v_z R^2 \frac{\partial T}{\partial z}}{4\alpha}\right\}.$$

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These consist of pages:

189, Appendix VI - The Four Friends